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(NASA-CR-170019) GEOPHYSICAL STUDY OF THE
STRUCTURE AND PROCESSES OF THE CONTINENTAL
CONVERGENCE ZONES: ALPINE-HIMALAYAN BELT
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GEOPHYSICAL STUDY OF THE STRUCTURE AND PROCESSES
OF THE CONTINENTAL CONVERGENCE ZONES - ALPINE-HIMALAYAN BELT

NASA GRANT NAG5-41

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ABSTRACT

The work done under NASA Grant NAG5-41, "Geophysical Study of the Structure and Processes of the Continental Convergence Zones - Alpine-Himalayan Belt" is described. This covers studies of the structure of the continental collision zones using seismic and body waves, theoretical modelling of the thermal regime of the convergence processes, and studies of earthquake mechanisms and deformation aspects of the model.

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I. INTRODUCTION

The Alpine-Himalayan tectonic belt contains the best examples of the continent-continent plate convergence processes at various stages of evolution. The collision of the Indian Plate with Asia gives rise to the development of the Himalaya and Tibet. The collision of the Arabian Plate with Eurasia is responsible for the modern tectonics of Iran and Turkey. The Alps represent an older collision zone.

Under the NASA Grant (NAG5-41) we are studying the structure and evolution of the continental collision zones using geophysical data and theoretical modelling techniques. In this report progress made in the past year is covered. The report is divided into three sections: Crustal Structure, Thermal Models, and Seismicity, Stress and Deformation. Under these sections the work in progress is described in brief reports; newly submitted papers are included as full manuscripts; and for those papers in press that will appear soon, only the abstracts are included.

II. CRUSTAL STRUCTURES ALONG THE ALPINE-HIMALAYAN BELT

We are studying the crust and upper mantle structures in the various segments of the Alpine-Himalayan belt to determine the properties and processes of the continental convergence zone. We use seismic surface wave phase and group velocities, travel times of body waves, and travel-time residuals. In the four studies that follow, a comparison of structures of different regions along the Alpine-Himalayan belt as well as detailed studies of Tibet, Hindu-Kush and Turkey are described.

II.1 STRUCTURE AND SEISMIC PROPERTIES OF THE ALPINE-HIMALAYAN CONVERGENCE ZONE

Phase and group velocities of both Rayleigh and Love waves between the periods of 10 and 60 seconds have been measured along a number of paths in the Middle East and Southern Asia. These new data have been combined with other available measurements (Niazi, 1968; Knopoff and Fouda, 1975; Bird, 1976; Bird and Toksöz, 1977). Some of the paths which have been used are shown in Figure 1. Coverage of the Middle East and South Central Asia is very good.

When possible, stations and events used for velocity measurements were chosen so that they all lie along the same great circle path. This allows the calculation of interstation velocities. Velocities were determined using a method developed by Taylor and Toksöz (1980). This method is a modification of that of Dziewonski et al. (1969) who used the cross-correlogram as an approximation to the interstation transfer function. The method of Taylor and Toksöz (1980) finds the transfer function using Wiener deconvolution. This transfer function represents the signal which would be observed at the second station given an impulse source at the first station. It contains information about the phase and group delay of the waves traveling between the stations. This method is particularly useful for closely spaced stations. In this case the use of the cross-correlogram can lead to errors in group velocity determinations.

A comparison of group velocities for different regions is given in Figure 2. Wide variations are evident. The velocities in Tibet are markedly different and much lower. Paths across the Arabian Peninsula have the highest velocities, although they are lower than the Canadian Shield velocities. It is interesting to note the significant variations of velocities that can be seen within a given region such as Iran.

Phase and group velocity data in Iran are shown in Figure 3. The velocities along the Zagros are noticeably different from those across the Iranian Plateau. Along the Zagros the airy phase is observed at longer periods, indicating a thicker crust than under the plateau. At periods greater than 35 seconds the velocities along the Zagros become higher than those across the plateau. This is most likely due as much to low velocities in the lower crust-upper mantle under the plateau as to high velocities in the Zagros.

A study of P_n waves supports these conclusions. A low P_n velocity of 7.9 km/sec was determined for waves crossing the Iranian Plateau. Chen et al. (1980) also found low P_n velocities (8.0 km/sec) in Iran, with no evidence for S_n propagation across the plateau (Chen, 1979, personal communication). P_n velocities are low (7.9 km/sec) under the Eastern Turkish Plateau also (Canitez and Toksöz, 1980). Velocities are higher (8.1 km/sec) under the Zagros, however, and the crust is thicker - about 50 km (Islami, 1972; Akascheh and Nasser, 1972).

A maximum likelihood technique was used to invert for velocity structures. Phase and group velocities were inverted simultaneously for shear velocity. The inversion is weighted in both model and data space. Crustal thickness was determined to be 45-50 km in Iran. Chen et al. (1980) suggest a crustal thickness of 34 km in Northern Iran and 49 km in the south. The surface wave data were for the path SHI-MSH. MSH is in northeastern Iran and SHI is to the southwest of the Zagros so the derived structure gives a good average for Iran.

Figure 4 shows observed (Patton, 1978; Chen and Molnar, 1981) and theoretical phase and group velocities for Tibet. The velocity profiles are shown in Figure 5. These indicate a 70 km thick crust displaying a low velocity zone. The resolution of the surface wave data is not sufficient to determine the presence of the low velocity layer, however, the velocity model is based on refraction studies which indicate a low velocity zone in the crust (Teng et al., 1980).

Velocities determined for the Arabian Peninsula are shown in Figure 6. Velocities from Knopoff and Fouda (1975) and Niazi (1968) are shown also. Phase velocities for the Canadian Shield (Brune and Dorman, 1963) are given for comparison. Figure 6 demonstrates systematic variations in velocities through this region. Both phase and group velocities tend towards higher velocities to the north and to the west. The north-south variation can be seen by comparing velocities along the paths AAE-SHI, Red Sea-SHI, SHI-HLW, and SHI-JER. The velocities

increase steadily as the path moves northward. An increase in velocity from east to west can be seen by comparing the curves for the paths Arabian Sea-JER and Gulf of Aden-JER. The differences decrease as period increases, indicating a more uniform structure at depth. The path from the Red Sea to SHI crosses the central region of the peninsula so it can be taken as an average structure for Arabia. An inversion of the velocities along this path gives a crustal thickness of about 38 km.

Figure 7 gives a comparison of shear velocity structures determined for three regions. The crustal thicknesses are 38 km, 45 km, and 70 km for Arabia, Iran, and Tibet, respectively.

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Figure Captions

Figure 1 - Map showing some of the paths crossing the Alpine-Himalayan convergence zone for which surface wave velocities have been measured.

Figure 2 - Rayleigh wave group velocities along various paths in South Central Asia.

Figure 3 - Rayleigh wave phase and group velocities in Iran.

Figure 4 - Rayleigh wave phase velocity (c) and group velocity (u) for the average of all paths crossing Tibet. Data from Bird (1979) and Chen and Molnar (1980).

Figure 5 - Crustal S-wave velocity model resulting from the inversion of Rayleigh wave data combined with the refraction P-wave profile (Teng et al., 1980) shown at right. Dotted area indicates the region of high attenuation.

Figure 6 - Rayleigh wave phase and group velocities across the Arabian Peninsula.

Figure 7 - Comparison of crustal structures in different regions. Heavy lines indicate the bottom of the crust (Moho discontinuity). The number in each layer is the shear velocity. "Iran" is the average model for the Iranian Plateau. The Arabian Peninsula (a shield) model is included for comparison.

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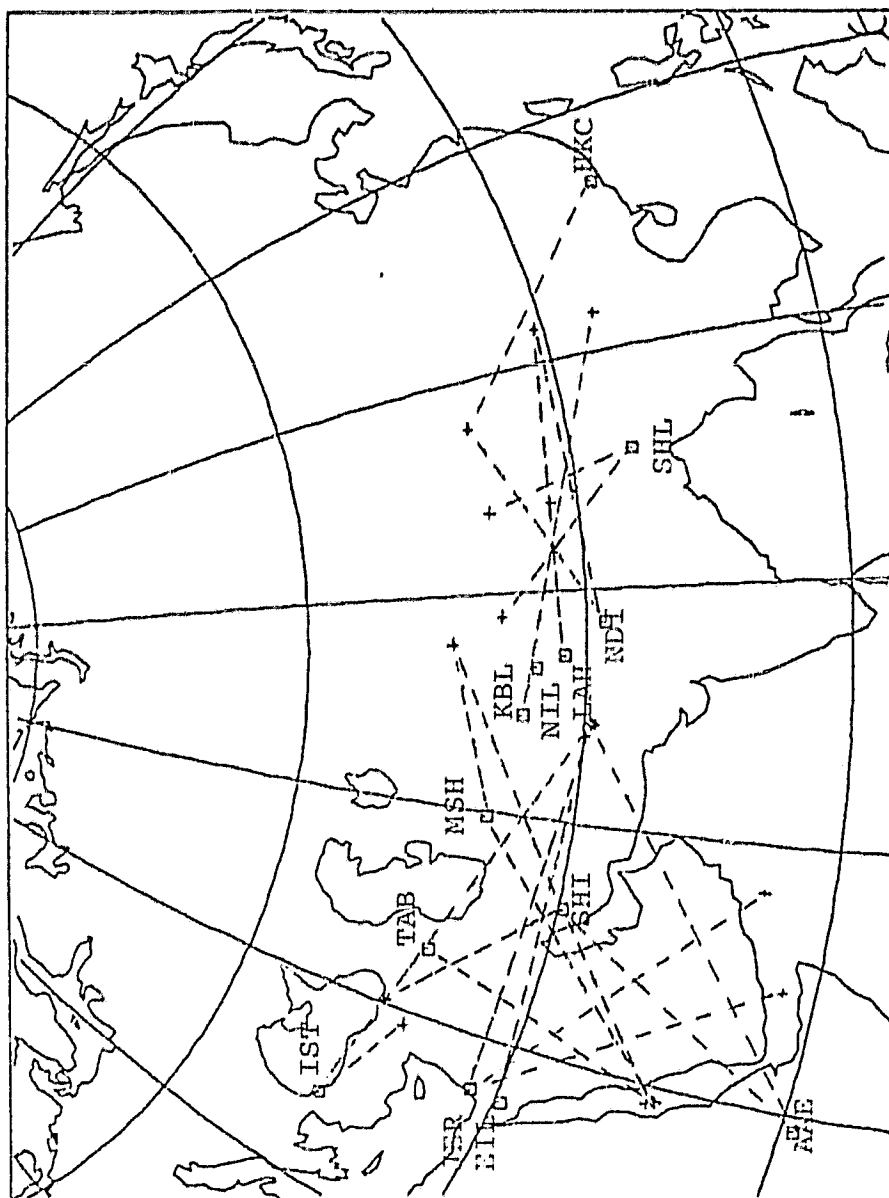


Figure 1

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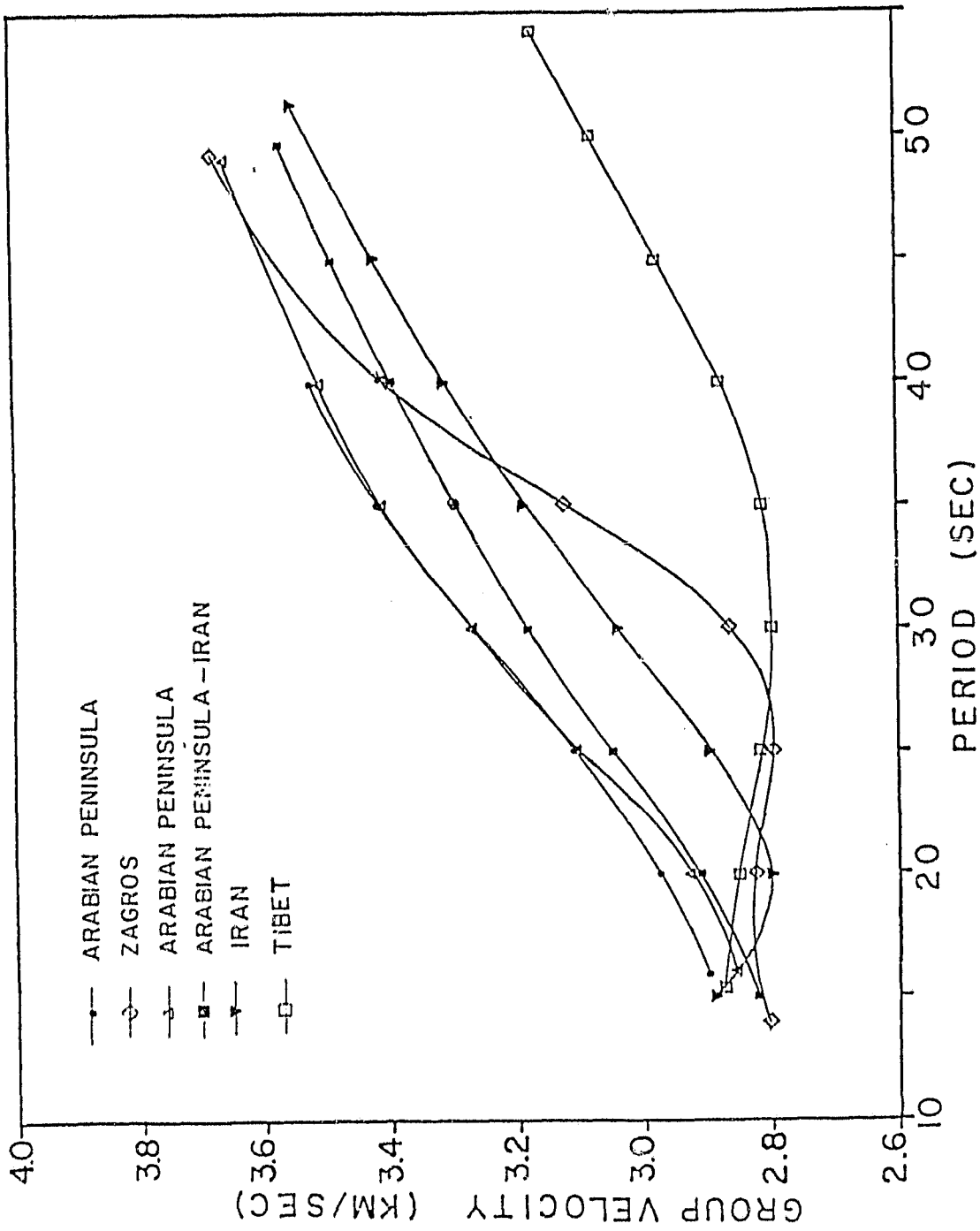


Figure 2

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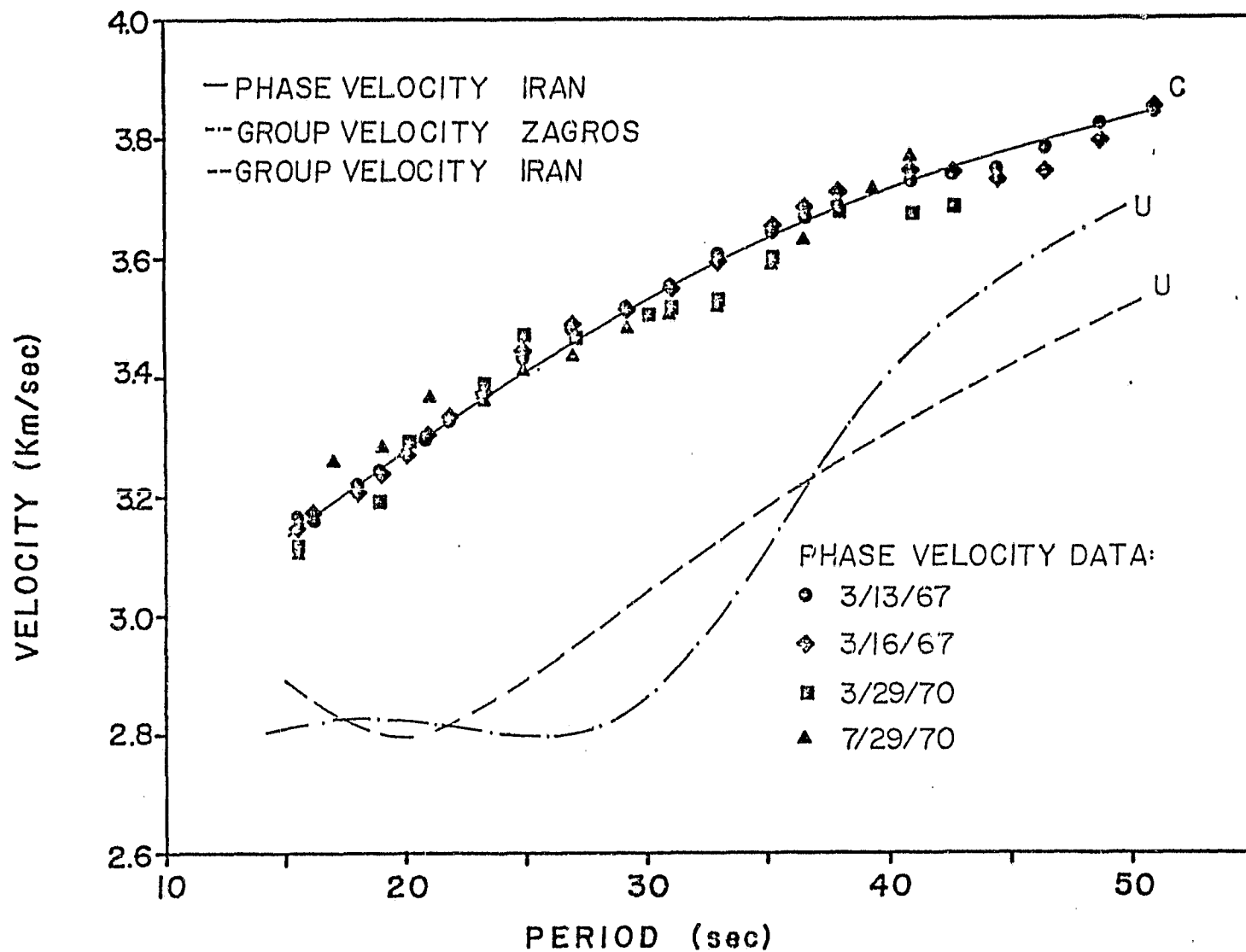


Figure 3

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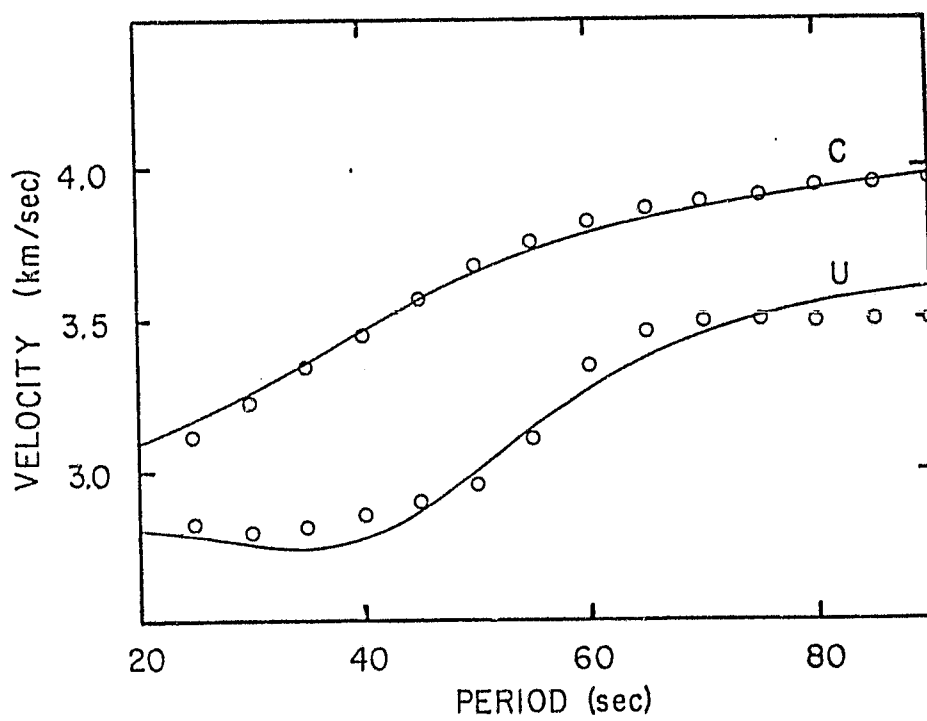


Figure 4

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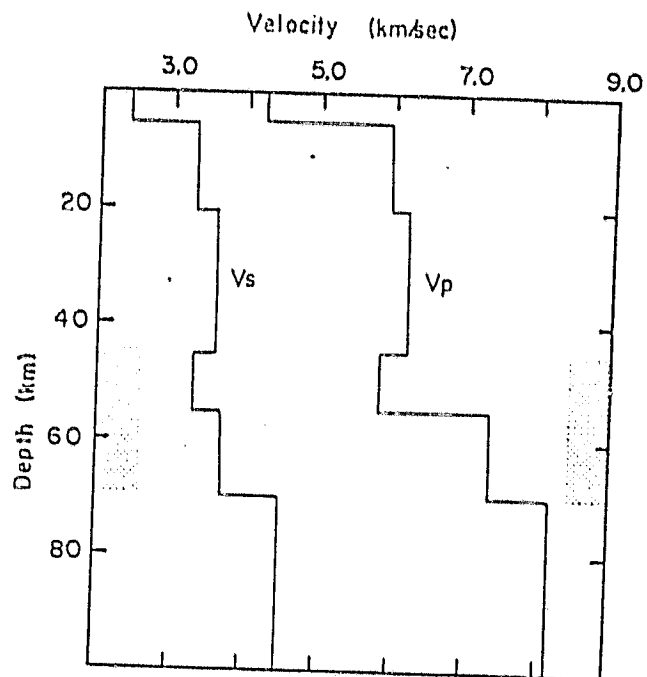


Figure 5

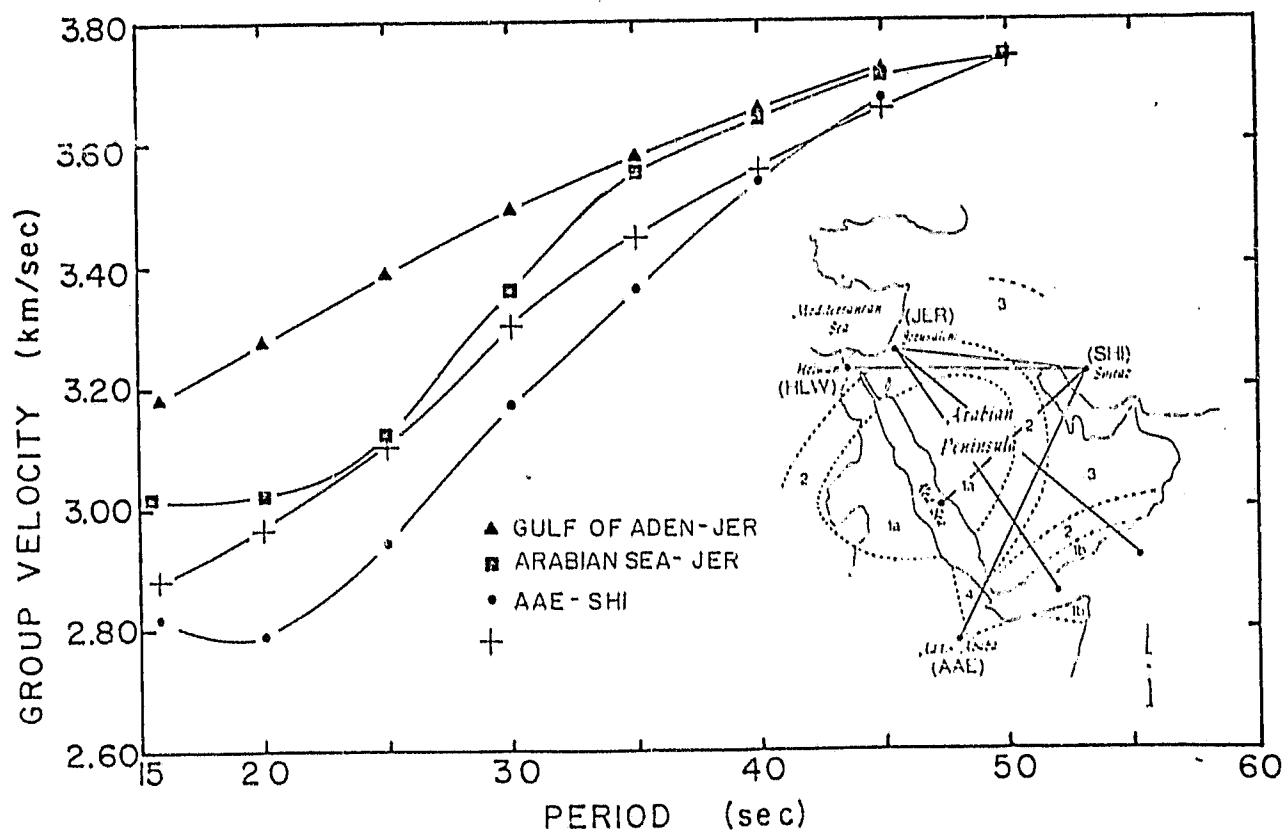
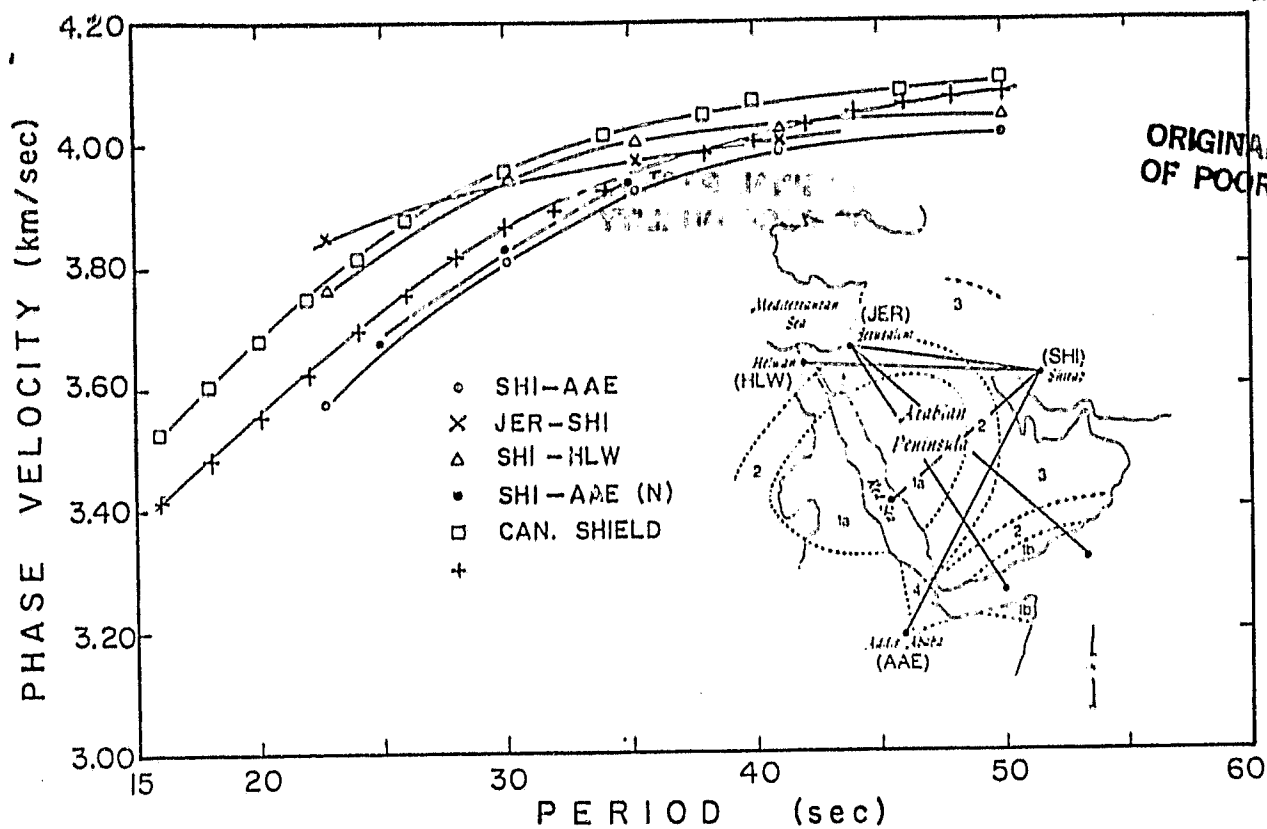


Figure 6

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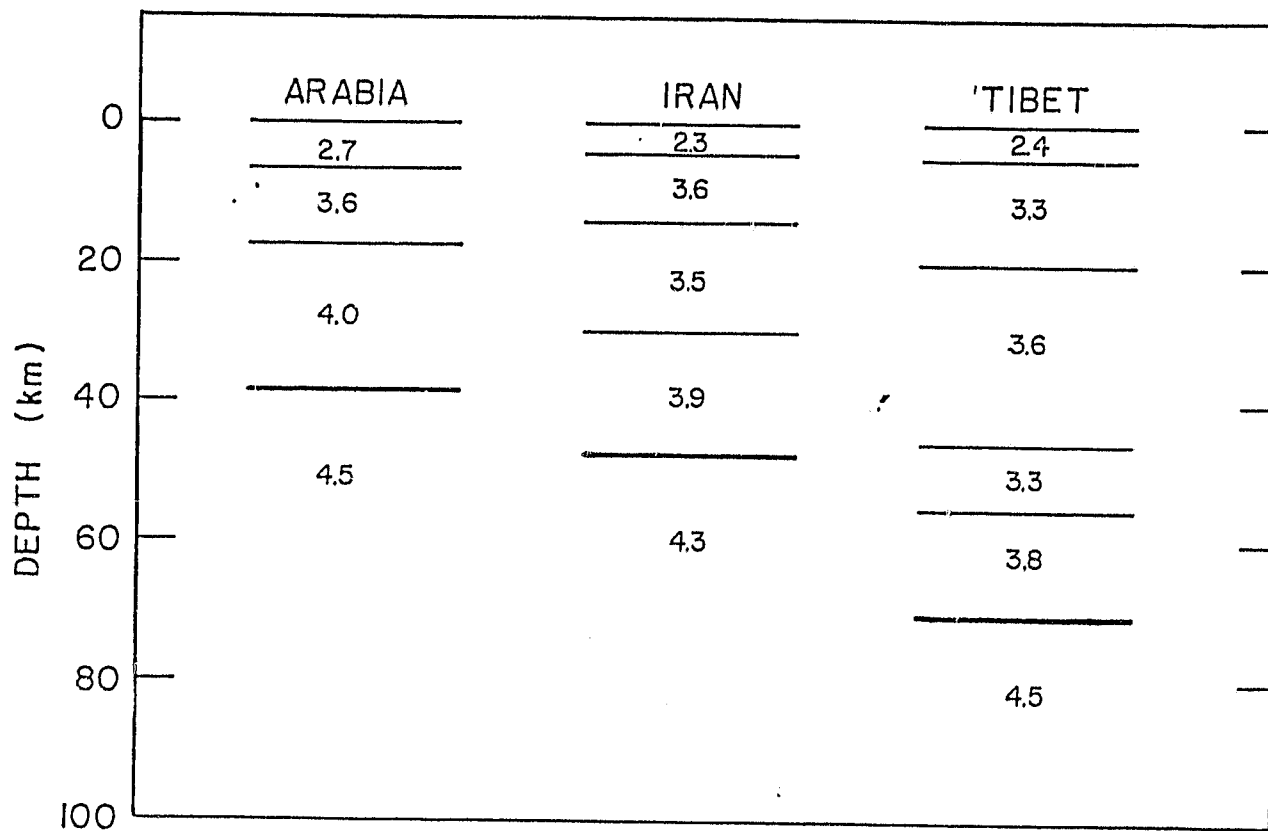


Figure 7

II.2 CRUSTAL STRUCTURE IN TURKEY*

Nezihi Canitez and M. Nafi Toksöz

ABSTRACT

Seismic travel times of P-waves and station residuals are used to study crustal structure and its lateral variations in Turkey. These data are complemented by the available phase and group velocities of surface waves and gravity Bouguer anomalies. The average P-wave station residuals are negative (early) in western and positive (late) in eastern Turkey. Local travel times give a P_n velocity of 7.9 km/sec in the east and 8.1 km/sec in the west. The travel time residuals are correlated with the negative Bouguer anomalies.

Combined with surface wave information, these data suggest that average crustal thickness is about 35 km in western Turkey (Aegean region), 30 km in northwestern Turkey, and greater than 40 km (most likely about 45 km) in eastern Turkey. Along the Black Sea coast the crustal thickness appears to be less than 30 km, but the travel time residuals are positive.

*Tectonophysics, submitted, 1980.

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INTRODUCTION

Turkey is a tectonically complicated area where several plates and continental fragments interact. To understand the surface geology and its evolution, it is important to know the crust and upper mantle properties. In this paper we study the crustal structure and its variation, using the available seismic travel times and station residuals.

The average elevation of Turkey is over 1000 m and it increases eastward to about 1900 m. The central part of the country is occupied by the Anatolide/Tauride platform. North of the platform the Pontic Mountains follow the Black Sea coast. In the south, the Taurus ranges follow the Mediterranean coast. In western Turkey, mountain chains are perpendicular to the Aegean coast and are separated by low plains.

The tectonics of Turkey is quite complicated. As shown in Figure 1, several tectonic units have been identified (Ketin, 1968). These are;

- (1) Pontide tectonic unit in northern Turkey including the region of Marmara
- (2) Anatolide tectonic unit making up central Anatolia
- (3) Tauride tectonic unit along the Mediterranean
- (4) Border folds in southeast Turkey.

Western Turkey consists of a set of horst-graben structures with general E-W orientation. The North Anatolian Transform Fault cuts both Pontide and Anatolide units.

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The left-handed East Anatolian Fault forms a triple-junction with the North Anatolian fault in eastern Turkey. East of the junction is characterized as an accretionary complex by Şengör and Yilmaz (1980). In western Turkey, an extensional tectonic regime is dominant.

Very little is known about the crustal structure and upper mantle properties in Turkey. The morphological and tectonic aspects of western Turkey resemble the Basin and Range province of the United States, yet it is not known whether the crust is thin and upper mantle velocities are low. Nor do we know if there are crustal thickness changes across major faults. The current information about the crust comes from gravity data (Canitez, 1962; Özelçi, 1973), dispersion of surface waves (Canitez, 1975; Ezen, 1979; Kenar and Toksöz, 1980), inversion of body-wave spectra (Kenar, 1978), magneto-telluric sounding (İlkışık, 1980) and one quarry blast refraction experiment (Üçer, Gürbüz, and Özdemir, 1980, personal communication). These data suggest a thicker crust in the east than the west and a rapid thinning of the crust toward the Black Sea north of the North Anatolian Fault Zone.

In this study we will use the teleseismic P-wave residuals to determine regional variations of crust/upper mantle structure. We will use the local travel time curves to establish upper mantle P_n velocities. Finally, we will combine these with the other data to present a model of crustal variation.

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Teleseismic P-Wave Residuals in Turkey

Teleseismic P-wave residuals are generally used to investigate average vertical properties of crust/upper mantle structure (Sengupta and Julian, 1976; Iyer, 1974; Iyer and Healy, 1972). In recent years, travel time residuals have been used for mapping the three-dimensional velocity structure beneath seismic arrays (Aki et al., 1977; Ellsworth, 1977; Ellsworth and Koyanagi, 1977; Menke, 1977; Zandt, 1978; Taylor and Toksöz, 1979).

There is a limited network of seismic stations in Turkey. Distribution is uneven and stations are especially sparse in eastern Turkey. In this study we used P travel time residuals from 125 earthquakes in 18 stations. The travel times were taken from ISC Bulletins. Elevation and ellipticity corrections were applied to the data. The absolute travel time residuals were calculated using J-B tables. The source effects (due to mislocation, error in origin time, inhomogeneities in the crust and upper mantle near the source) are minimized by calculating "relative residuals" by subtracting average event residuals from the absolute residuals (see: Ellsworth, 1977; Zandt, 1978; Taylor and Toksöz, 1979). Average station residuals are calculated using the relative residuals. In our sign convention, negative residuals correspond to early arrivals and positive residuals to late arrivals, respectively. We chose earthquakes in the epicentral distance range 60° - 85° ,

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where the angles of incidence for P waves are between 16° and 22° .

We used more than 1000 residuals from 125 events to calculate the average station residuals. To test the stability, we compared the average station residuals determined using a subset of 50 events, with those of 125 events. There was no significant change. Two typical examples of the distribution of teleseismic

P-wave residuals at seismic stations are given in Figure 2.

As is seen in the histograms, the distributions are well-behaved.

The average relative station residuals in Turkey are shown in Figure 3. Triangles correspond to negative (early), open circles to positive (late) arrivals. As is seen in the figure, distribution of residuals is systematic. In western Turkey and the region of the Marmara Sea, residuals are predominantly negative. These correspond to early arrivals. In northern and eastern Turkey, the relative residuals are all positive meaning late arrivals.

Sengupta and Julian (1976) calculated absolute station residuals using deep-focus earthquakes. Although they used only 5-6 events, their results (CIN: - 0.90; ISK: -0.62; KAS: 0.86) are consistent with our relative average residuals. Dziewonski (1979) calculated the station anomalies for P-wave travel times for 751 stations all over the world. His absolute residuals are also consistent with ours with a linear correlation coefficient of 0.83.

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The distribution of the residuals in Figure 3 is informative. The negative residuals in western Turkey are surprising. Looking at the general tectonics of the area, one would have expected positive residuals and lower velocities. We will discuss this region further utilizing all data.

The residuals in northern and eastern Turkey are all positive. The positive residuals in the east are consistent with the geological evidence indicating a crustal thickening in that area (Şengör and Kidd, 1979). The late arrivals at the KAS, KVT and MGN stations are unexpected results because of the thinning of the crust toward the Black Sea (Canitez, 1962; Kenar, 1978; İlkışık, 1980). It is clear that the 0.7 sec time delay compared to the west could not be explained only with the physical and structural changes within the crust; the whole lithosphere might be responsible for this.

In the areas showing lateral changes in crust/upper mantle structure, a linear correlation between travel time residuals and gravity anomalies is expected (see: Taylor and Toksöz, 1979; Fletcher et al., 1978). Figure 4 shows the correlation between our relative P-wave residuals and Bouguer gravity values, read from the map of Özelçi (1973). There is a reasonable correlation between the two data sets except for the KAS and MGN stations in northern Turkey, again indicating some unusual characteristics of crust/upper mantle structure between the North Anatolian Fault Zone and the Black Sea.

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P_n velocities

P-wave travel times up to 10° epicentral distance as reported in the ISC Bulletins have been used to investigate upper mantle velocities in Turkey. Although earthquake travel time data are affected by individual station reading errors, errors in source location, origin time and focal depth, for large data sets errors are generally averaged. We used the data from 36 earthquakes and 18 stations. For western Turkey, we used earthquakes from southwestern Turkey and the southern Aegean Sea. For eastern Turkey we used the earthquakes from eastern Turkey and near the Turkish/Iranian border. Earthquake hypocenters and two examples of the ray paths are shown in Figure 5 to illustrate the coverage. The travel time curve of waves in western Turkey is shown in Figure 6. A least squares approximation of 180 data points gives a P_n velocity of 8.1 km/sec. This result is consistent with the 8.13 km/sec P_n velocity obtained from a quarry blast experiment in the Adapazari region (Üçer, Gürbüz, and Özdemir, 1980, personal communication). The same velocity has been reported for the Mediterranean Sea and for the natural prolongation of western Turkey toward the Aegean Sea basin (Makris, 1975, 1978; Makris and Vees, 1977). The result obtained here is consistent with the early arrivals of teleseismic P-waves in western Turkey.

The travel time curve for eastern and northern stations is shown in Figure 7. A least squares approximation of these

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data gives a P_n velocity of 7.9 km/sec. The relatively low upper mantle velocity obtained is consistent with the late positive station residuals.

Discussion and Results

Two important results have been obtained in the present study. First, teleseismic P-wave residuals are negative in the west and positive in the east. The time delay compared to the west is about 0.7 sec. Second, there is a 2% difference in upper mantle velocities (8.1 km/sec in the west vs 7.9 km/sec in the east) between the two regions. These results indicate that there are some major differences in crust/upper mantle structure under different regions of Turkey. Using these results along with all other available data, we will synthesize a generalized model of the crustal structure and its variation.

Crustal structure in the Mediterranean and Aegean Seas adjacent to western Turkey has been investigated in some detail. The data presented by Makris (1975, 1976, 1978), Makris and Veis (1977), Jongsma (1974, 1975) and Morelli et al. (1975) indicate that the eastern Aegean Sea basin is the natural prolongation of western Anatolia. Makris (1976, 1978) gives 35 km crustal thickness along the western coast of Turkey.

Bouguer gravity decreases eastwards with a smooth gradient (Üzelçi, 1973). This indicates a slight increase in crustal thickness. From surface wave dispersion Canitez (1975) suggested

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a 36 km average crustal thickness for western Turkey. Starting with the North Anatolian Fault Zone, the north-south gradient of the gravity field becomes very high. Gravity field measured along three profiles across the western part of the North Anatolian Fault represents a -120mgal Bouguer anomaly (Canitez, 1962). This indicates a rapid change in crustal thickness in the north-south direction. Using these data along with surface wave dispersion, Canitez (1962) argued that the 36 km average thickness of the crust decreases towards the Black Sea, and is about 28 km at the coast. Local crustal structure near Istanbul has been studied by Kenar (1978). From the inversion of body-wave spectra, he found a thickness of 28-30 km for the Istanbul region. Qualitative interpretation of the group velocities of surface waves recorded in Istanbul has been made by Ezen (1979). He argued that the interferences seen in surface waves travelling through the Aegean Sea and western Turkey are due to the lateral inhomogeneities in the earth's crust. A magneto-telluric survey has been undertaken by Ilkışık (1980) in Thrace, northwestern Turkey, at nine localities along a north-south profile 80 km long between Kırklareli and Tekirdag. He found that the crustal thickness is about 40-44 km at Kefcaz Yolu on the Istranca Massif in the north, and thins to about 28-32 km towards the central part of the basin and then thickens to about 32-35 km at the southernmost part of the profile. Only one

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refraction experiment has been done near Adapazari in the Marmara region of Turkey (Üçer, Gürbüz and Özdemir, 1980, personal communication). They recorded a quarry blast at 13 stations along a profile 144 km long between Adapazari and Fındıkköy. They suggest a 28-30 km thick crust with $V_g = 5.37$ km/sec and $V_n = 8.13$ km/sec.

If we summarize the results for western Turkey directly to east of the Aegean Sea, the crust is about 35 km thick, P_n velocity is 8.1 km/sec and station residuals are negative. These properties are typical for an average tectonic continental crust. Yet, western Turkey is tectonically active. East-west oriented grabens and earthquake mechanism (McKenzie, 1978; Canitez and Üçer, 1967; Dewey and Şengör, 1979; Alptekin, 1973) suggest a north-south extension. One geologic interpretation suggests this area may have undergone 30-50% extension (Şengör, 1978). Although there are hot-springs in many parts and some recent volcanics, especially associated with the Gediz and Simav grabens (Kulaits), there is no direct evidence indicating a broad regional heating of the lithosphere.

For eastern Turkey, P_n velocity is 7.9 km/sec. Station travel time residuals are positive. There is a broad region of negative Bouguer anomalies (Özelçi, 1973). The average crustal thickness obtained from Love and Rayleigh wave dispersion for paths extending from western Iran to Istanbul is about

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40-43 km (Kenar and Toksöz, 1980). This suggests that in eastern Turkey, the crust is thicker than 40 km and most likely it is about 45 km. The thick crust is consistent with the models of geologic evolution of the region (Sengör and Kidd, 1979). It is also similar to other high plateaus (Iranian Plateau, Tibet) in that the region is characterized by low P_n and S_n velocities and high attenuation (Toksöz and Bird, 1977). It is reasonable to assume that temperatures in the lower crust and uppermost mantle are higher than typical continental values.

The region between the North Anatolian Fault Zone and the Black Sea poses a problem. Both gravity and seismic data indicate a relatively thin (about 28-30 km) crust. Yet the travel time residuals are consistently positive at stations MGN, KAS, KVT and TBZ. The accretional history of this region is complicated. Positive residuals may be due to lower velocities in the asthenosphere, petrology of the crust and lithosphere (i.e., rich ultramafic composition) or a combination of effects.

The regional variations of crustal thickness in Turkey are summarized in Figure 8. The boundaries and details of crustal thickness changes are not well established. The map serves as a generalized guide, for regional assessment and for future studies.

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Figure Captions

Figure 1. Map of Anatolia and surrounding regions showing the active plate boundaries and the tectonic subdivisions (compiled by Şengör, 1979).

Figure 2. Distribution of P-wave residuals at ISK (Istanbul-Kandilli) and KAS (Kastamonu). Class interval is 0.6 sec.

Figure 3. Distribution of teleseismic P-wave residuals in Turkey. Triangles show negative (early), open circles positive (late) arrivals. The figure for RAM station is the absolute residual and is taken from Dziewonski (1979).

Figure 4. Bouguer gravities vs teleseismic P-wave residuals. Gravity values are read from Ozelci's (1973) map. Notice the negative correlation between gravity field and travel time anomalies. Scattering might be mostly due to the effect of surface geology. Notice also abnormal positive residuals at KAS and MGN stations of northern Turkey.

Figure 5. Locations of seismic stations, epicenters and two examples of ray paths used for local travel times for western and eastern Turkey.

Figure 6. P_n travel time curve for western Turkey. Least squares approximation gives 8.1 km/sec.

Figure 7. P_n travel time curve for eastern Turkey. Least squares approximation gives 7.9 km/sec.

Figure 8. Variation of the crustal thickness in Turkey.

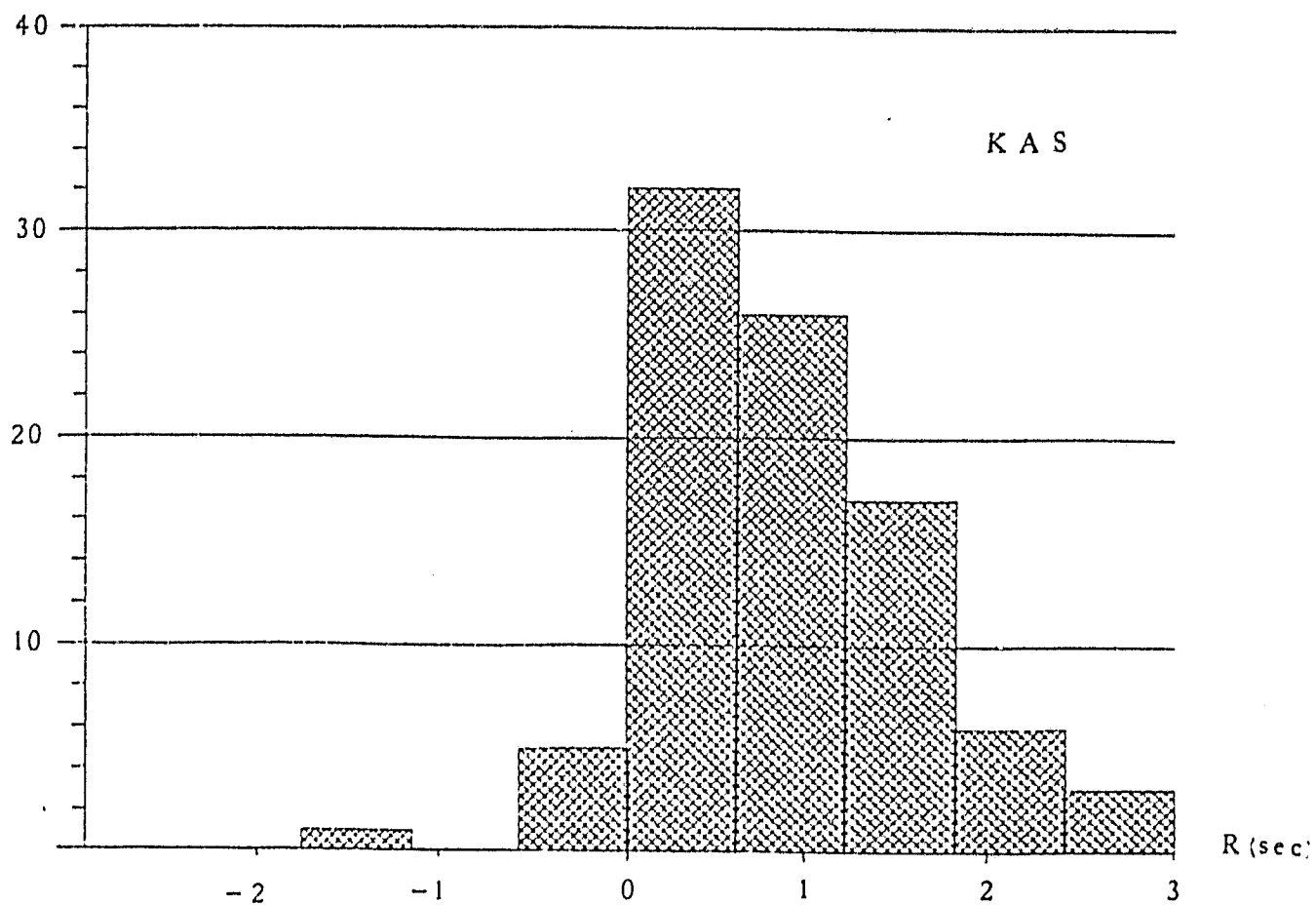
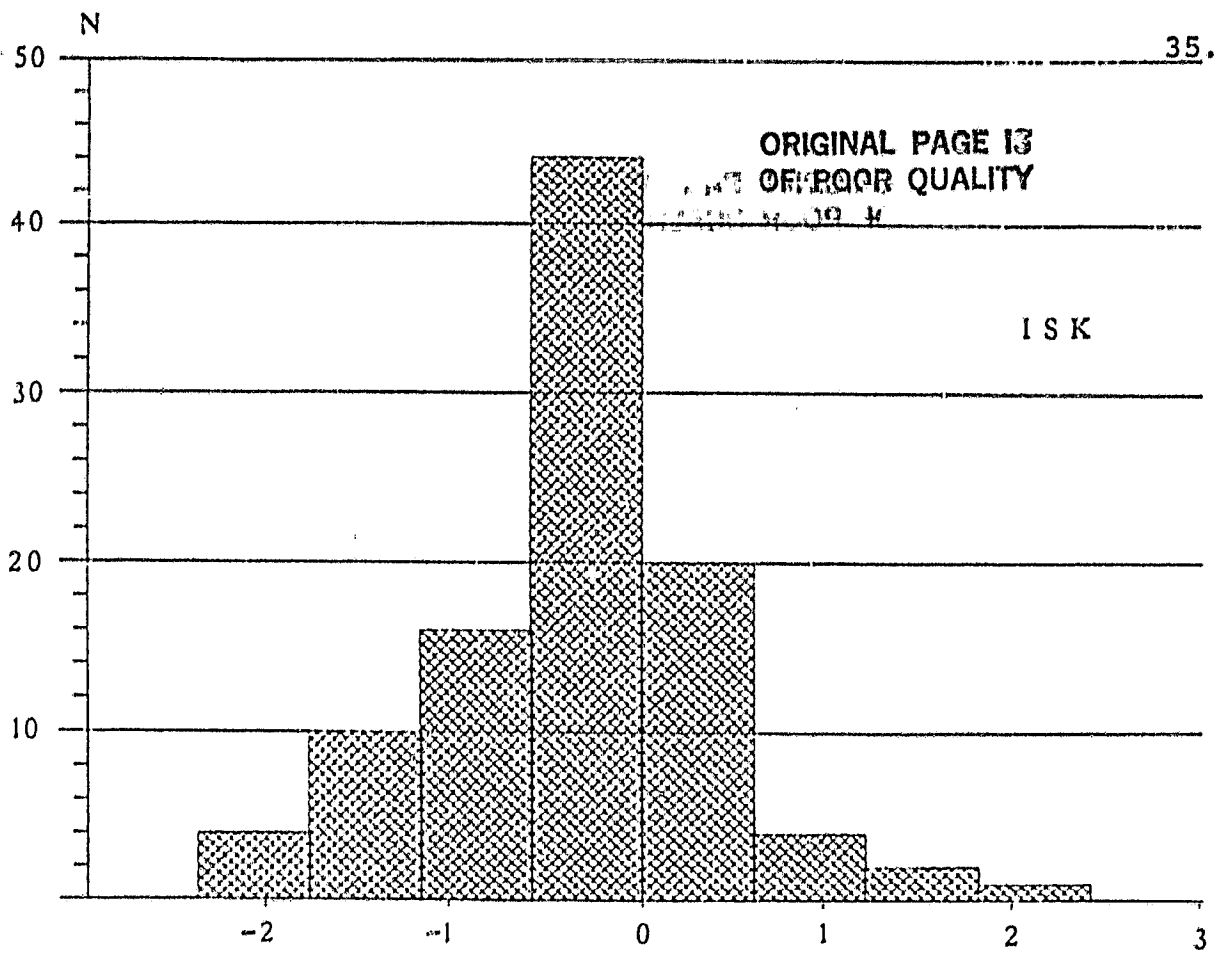


Fig. 2

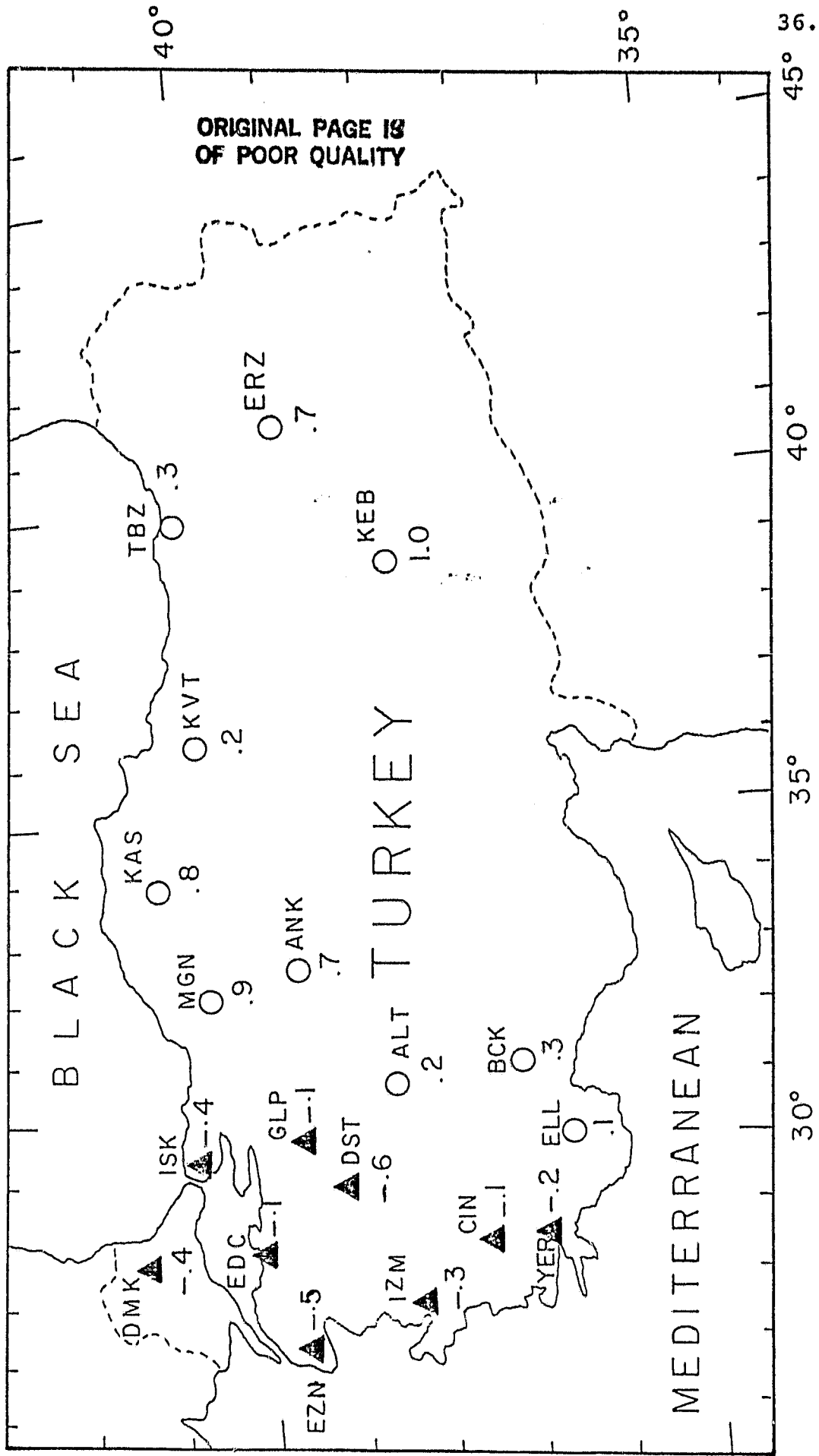


Fig. 3

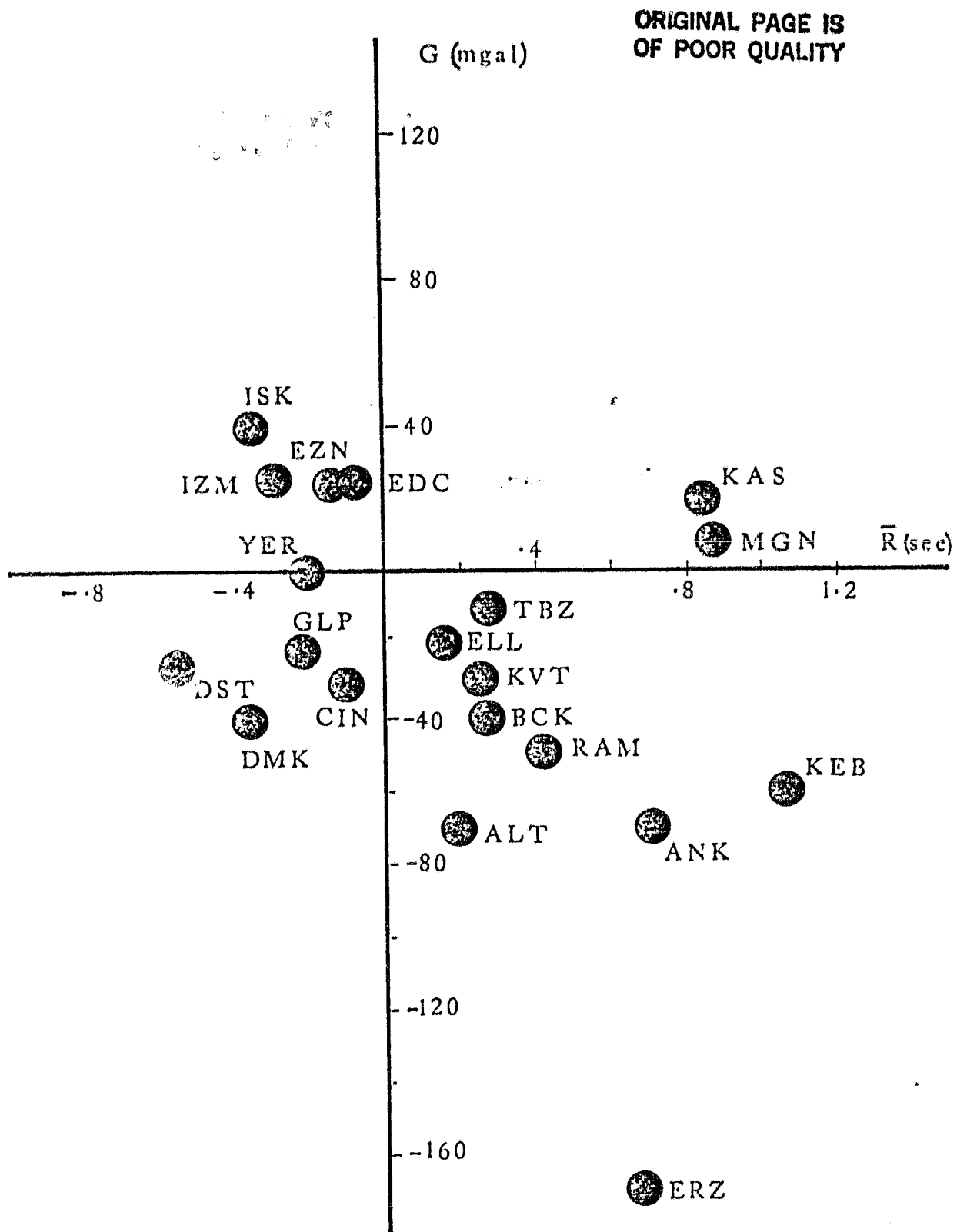


Fig. 4

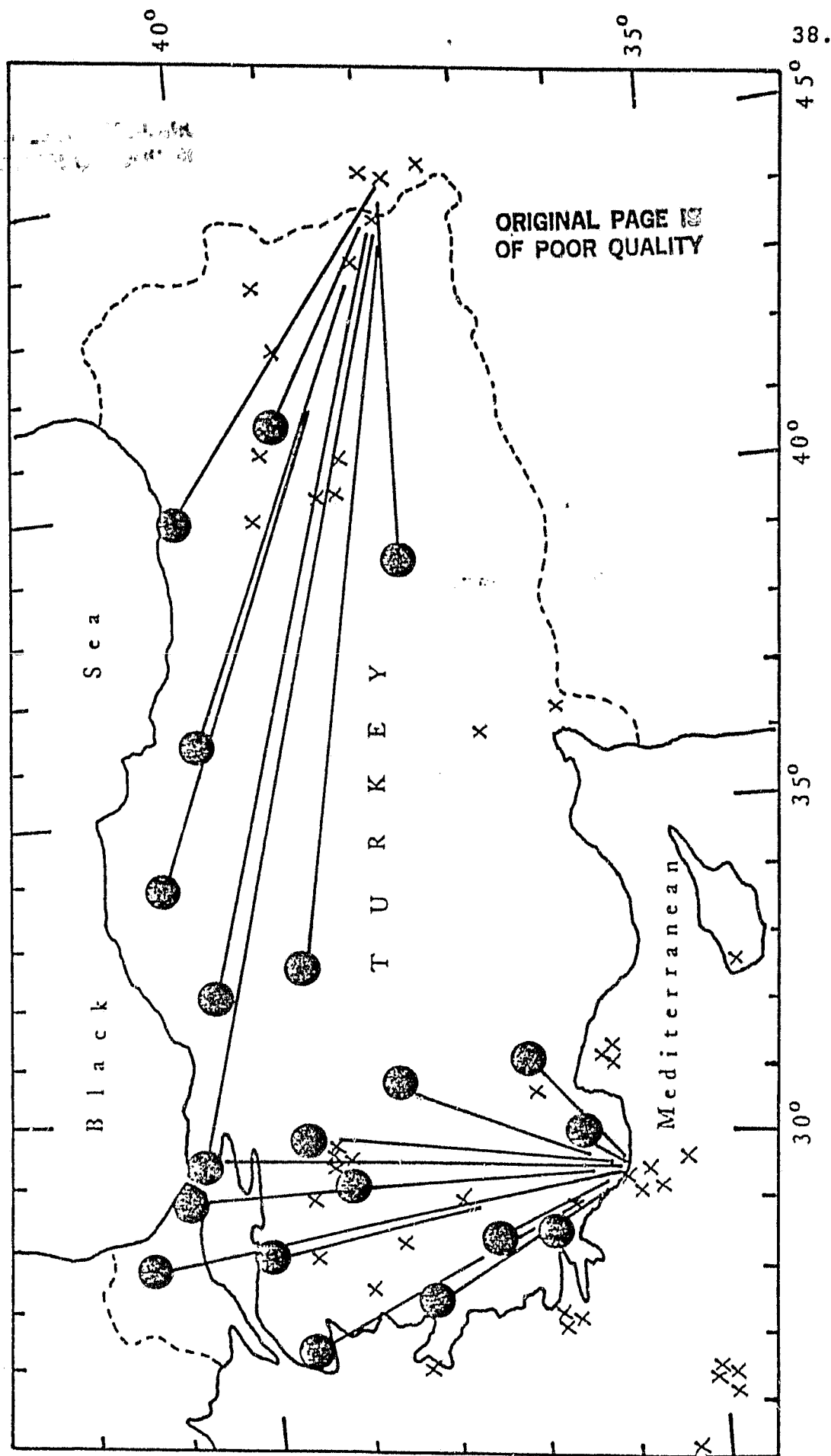


Fig. 5

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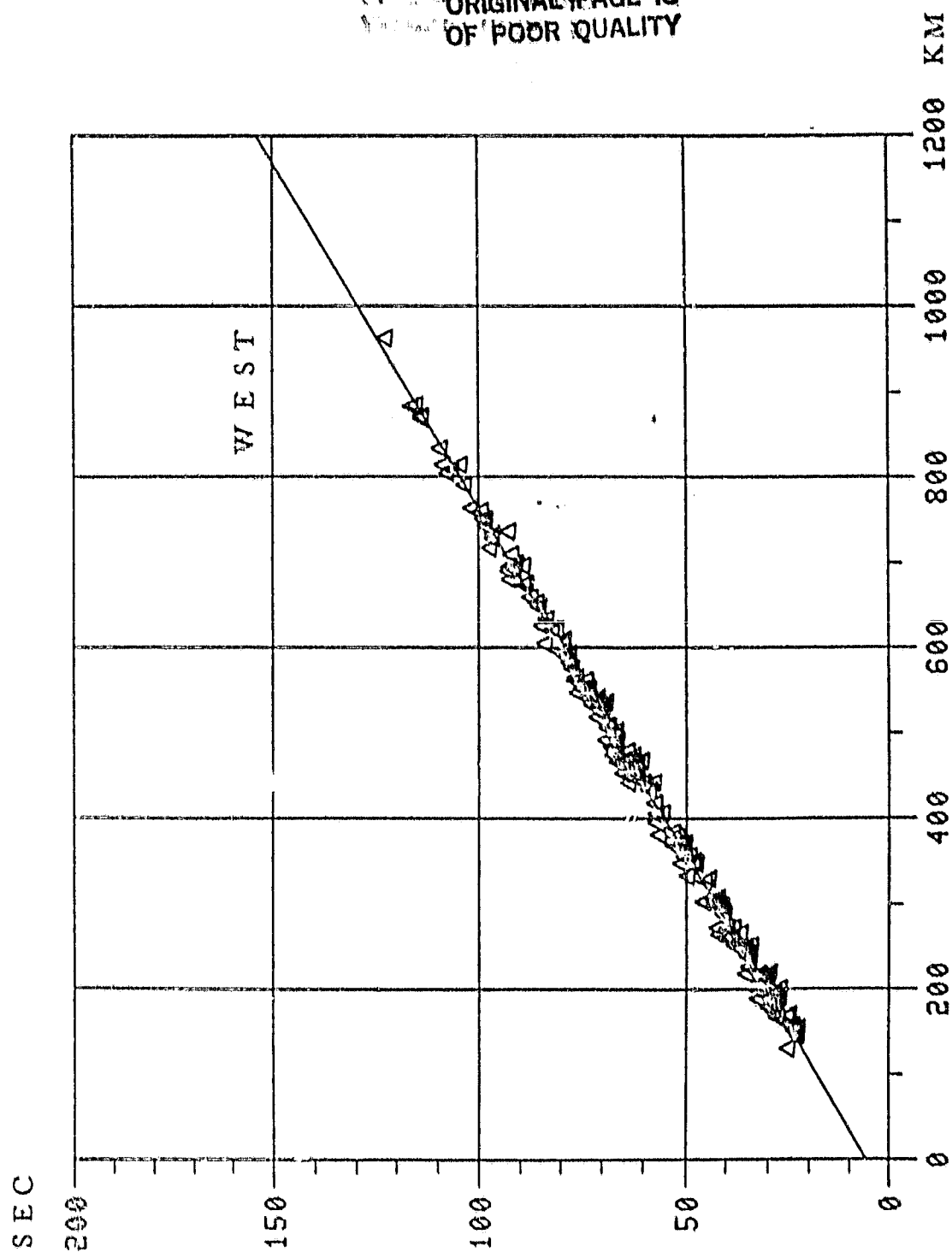


Fig. 6

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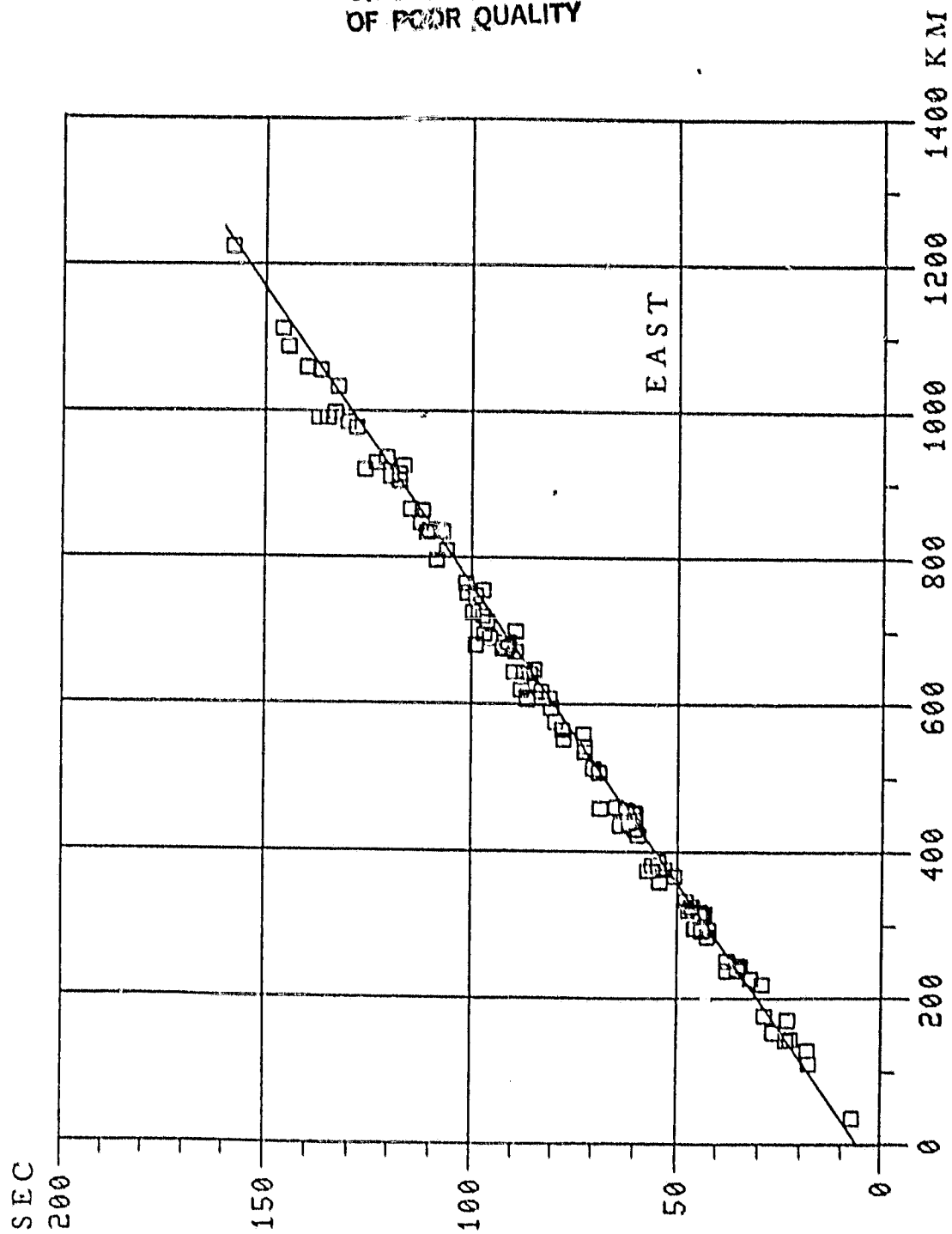


Fig. 7

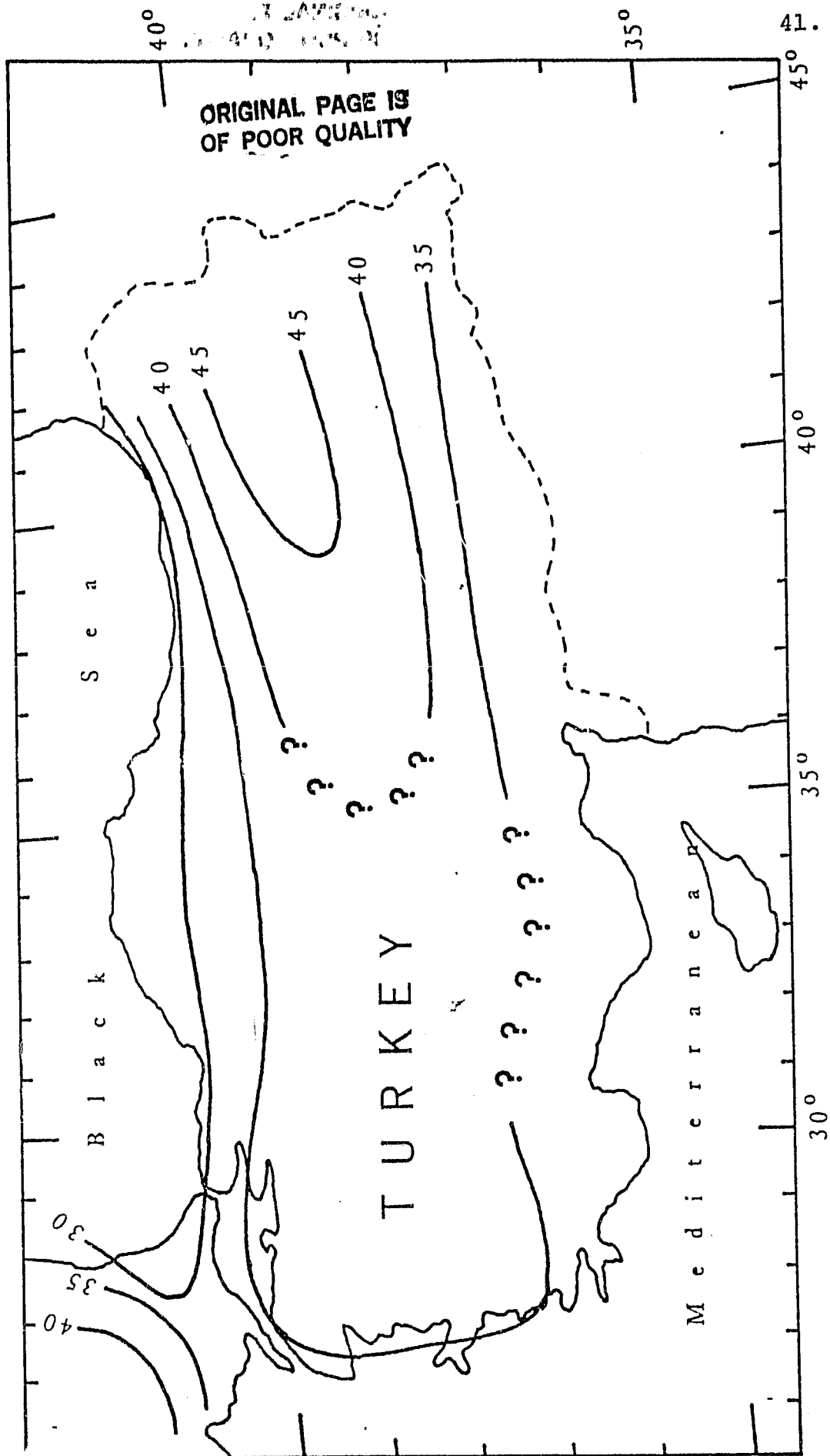


Fig. 8

II.3 CONSTRAINTS ON THE SEISMIC WAVE VELOCITY STRUCTURE
BENEATH THE TIBETAN PLATEAU AND THEIR TECTONIC IMPLICATIONS*

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Abstract. We combine observations of group and phase velocity dispersion of Rayleigh waves, of the waveform of a long-period P_L phase, of P_n and S_n velocities from unreversed refraction profiles using earthquakes, and of teleseismic S-P travel time residuals to place bounds on the seismic wave velocity structure of the crust and upper mantle under Tibet. From surface wave measurements alone, the Tibetan crustal thickness can be from 55 to 85 km, with corresponding uppermost mantle shear wave velocities of about 4.4 to 4.9 km/s, respectively. The P_n and S_n velocities were determined to be 8.12 ± 0.06 and 4.8 ± 0.1 km/s, respectively, using travel time data at Lhasa from earthquakes in and on the margins of Tibet. Combining these results, the crustal thickness is most likely to be between 65-80 km with an average shear wave velocity in the upper crust less than 3.5 km/s. A synthesis of one P_L waveform does not provide an additional constraint on the velocity structure but is compatible with the range of models given above. In contrast to observations obtained for eight earthquakes in the Himalaya, measurements of both teleseismic S and P wave arrival times for nine earthquakes within Tibet show unusually large intervals between P and S compared with the Jeffreys-Bullen Tables. Thus the P_n and S_n velocities apparently do not reflect high velocities in the mantle to a great depth beneath Tibet. From the dependence of the seismic velocities of olivine on pressure and temperature and from the similarity of the measured P_n and S_n velocities beneath Tibet and beneath shields and platforms, the velocities at the Moho beneath Tibet are compatible with the temperature being 250° - 300° higher than beneath shields and platforms, i.e., 750°C if the temperature beneath the platforms is close to 500°C . Such a temperature could reach or exceed the solidus of the lower crust. Simple one-dimensional heat conduction calculations suggest that the volcanic activity could be explained by the recovery of the geotherm maintained by a mantle heat flux of

*J. Geophys. Res., in press, 1981.

about 0.9 HFU at the base of the crust. If the distribution of radioactive heat production elements were not concentrated at the top of the crust, radioactive heating could also contribute significantly to the recovery of the geotherm and thus lower the required mantle heat flow. Thus the idea of a thickened crust in response to horizontal shortening is compatible both with these data and with these calculations.

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II.4 The Velocity Structure of the Pamir-Hindu Kush
Region: Possible Evidence of Subducted Crust*

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ABSTRACT

The arrival times of compressional (P) and shear (S) waves from approximately 580 microearthquakes recorded by a temporary array in the Pamir-Hindu Kush region in central Asia are used to deduce one- and three-dimensional velocity structures of this region. The results for one-dimensional structures imply that the Moho is at 70 ± 5 kilometers depth. Also, there is a velocity reversal near 160 kilometers depth, which is inferred to be the beginning of the low velocity zone. This reversal continues to depths of approximately 230 kilometers. Below 230 kilometers, velocities are somewhat higher than those of normal mantle at similar depths (9.3 km/sec vs. 8.4 km/sec for P waves). The outstanding feature of the results for three-dimensional velocity structures is a broad (>40 kilometers), centrally located region with 8% to 10% lower velocities than those in the surrounding regions. This low velocity region envelopes the seismic zone at depths between 70 and 150 kilometers. The region may actually extend beyond these depths, but the results for shallower and deeper structure lack sufficient

resolution to decide. Several tests, using both hypothetical and real data, were performed to estimate the reliability of the three-dimensional solutions. The results of these tests suggest that the inferred velocities are reasonably accurate representations of the average velocities in the blocks, although one must be cautious of the effects of averaging in interpreting the solution. The low velocity region is inferred to be a manifestation of substantial quantities of subducted continental crust. Therefore, while subduction has occurred in the Pamir-Hindu Kush, the results of the three-dimensional inversions suggest that continental, rather than only oceanic, lithosphere has been subducted to depths of at least 150 kilometers.

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III. THERMAL MODELS OF CONVERGENCE ZONES

The thermal regimes of plate convergence and subduction zones determine the physical properties of the lithosphere and deformation mechanisms associated with convergence. Present day temperatures can be estimated from heat flow, volcanic activity and melting curves and indirectly from seismic velocities and attenuation. The relationship of these to the processes in the lithosphere and asthenosphere are generally inferred from model calculations.

In this section we describe thermal models of convergence zones. The first paper deals with the thermal consequences of the crustal thickening in Tibet. In the following three papers, asthenospheric convection and the role of thermal boundaries on temperature regimes are discussed.

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III.1 CRUSTAL EVOLUTION AND THERMAL STATE OF TIBET*

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ABSTRACT

The crustal structure and seismic attenuation under Tibet have been studied by seismic body and surface waves. The crust is about 70 km thick. Average velocities in the crust are lower than average continental values. The attenuation in the lower crust (below about 50 km) is very high. The Rayleigh waves in the period range of 30-50 seconds are highly absorbed. Shear wave Q values may be as low as $Q_s = 10$ in the lowermost part of the crust. The above properties may be explained by the thickening of the crust through internal deformations and elevated temperatures. We have examined the thermal consequences of such an evolution. The initial conditions are those determined by the subduction processes that have preceded the Himalayan episode of continental collision. Internal heating due to redistribution of crustal radioactivity contributes significantly to the thermal regime. The effects of shear strain heating during crustal thickening are less significant. In a period of about 40 m.y., temperatures in the lower crust will exceed 700°C. This could product partial melting and result in excessive attenuation of shear waves.

INTRODUCTION

Tibet is the largest continental plateau on the earth. It covers approximately 700,000 km² with an average elevation of 5 km. Its formation and present state are in part tied to the collision of India with Eurasia. In this paper we propose a model for the evolution of the Tibetan Plateau based on geological and geophysical data and we test the thermal consequences of such an evolution using finite difference calculations.

Our model implies that before the continental collision Tibetan lithosphere was heated from underneath by induced mantle convection. After collision crustal thickening occurred by internal deformation; the heat flux from the mantle and radioactive heat sources in the crust raised the temperatures in the lower crust and caused partial melting. This model is consistent with the geophysical data. In the paper we first review these data and then describe the results of the thermal calculations.

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The data we consider in constructing our models come from gravity, magnetics and seismic data.

The best coverage of gravity data for Tibet comes from satellite data. Because of its large area and great elevation, Tibet stands out as a major feature on satellite gravity maps. In Fig. 1 the free air gravity map of Tibet and vicinity based on the GEM-8 model is shown⁽¹⁾. The Tibetan anomaly can be modeled by an isostatically compensated or partially compensated density distribution. Satellite data alone cannot reveal the degree of isostatic compensation since the anomalies in the mantle contribute strongly to the measured field. Limited land-based data^(2,3) favor compensated models. The great elevation of Tibet, if explained as the result of the isostatic compensation of the 70 km thick crust, requires a very low average crustal density $\rho_c \approx 2.8 \text{ g/cm}^3$. An alternate explanation⁽⁴⁾ is that only part is due to isostatic compensation and part is due to the compression of the area by the continued convergence of India and Asia.

Magnetic field measurements from satellites show a very large negative anomaly over Tibet (Fig. 1). Magnetic data only see crustal anomalies⁽⁵⁾. Inversion of the data in terms of crustal magnetization (Mayhew, 1980, personal communication) shows the entire plateau region to be a fairly uniform area of slightly lower than normal magnetization. The far western region of the plateau is a region of particularly low magnetization. A possible cause for the generally low level of total magnetization of the crust is that the depth to the Curie isotherm, taken to be about 500°C ⁽⁵⁾, is less than for normal crust. Recent calc-alkaline volcanics, which have been found over much of Tibet^(6,7,8), suggest elevated temperatures and possible partial melting in the crust. Abundant hot springs are found throughout southern Tibet and some active hydrothermal areas are found in northern Tibet, but many areas there now appear to be cooler than in the past⁽⁹⁾.

The most specific information about the crustal properties under Tibet comes from seismic data. Although there are relatively few earthquakes and stations in Tibet, seismic waves propagating across the Tibetan Plateau provide information about the average properties. Rayleigh waves from many paths crossing Tibet have been analyzed by a number of investigators to obtain phase and group velocity curves^(10,11,12,13,14). These data can be complemented by some body wave data⁽¹¹⁾, refraction results⁽¹⁵⁾, and attenuation measurements^(10,16).

In this study we inverted the phase and group velocity data, together with the refraction data, to obtain crustal structure and velocities. Some examples of Rayleigh waves crossing Tibet are shown in Fig. 2. The source is an explosion at Lop Nor. Note that as the fraction of the Tibetan segment increases in the total path, both dispersion and attenuation characteristics change. For example, Rayleigh waves recorded at Kabul (KBL) have well-defined long-period ($T \geq 40\text{s}$) waves but these are strongly attenuated across Tibetan paths leading to New Delhi (NDI) and Shillong (SHL). This suggests that strong seismic attenuation is taking place in the lower crust under Tibet. The attenuation as a function of period for the two paths from Lop Nor which cross Tibet is shown in Fig. 3.

The Rayleigh wave phase and group velocity data from all available paths are averaged and inverted to obtain a crustal velocity model (Tubman, 1980, personal communication). The data and theoretical fit are shown in Fig. 4. In the inversion, the compressional wave velocities were specified from the

refraction data for central Tibet⁽¹⁵⁾. Shear wave velocities, densities and crustal thickness were iterated to obtain the best fit.

The velocity profiles of the final model are shown in Fig. 5. Important features of the model are: (a) the crust is 70 km thick and this is verified independently by both the refraction and surface wave data; (b) the average P and S wave velocities in the crust are low; (c) there may be a "lower" velocity layer below a depth of 45 km, although surface waves alone can resolve neither the thickness nor the actual velocity of this layer; and (d) velocities below the Moho, although at 70 km depth, correspond to typical continental upper mantle velocities ($V_p = 8.15$ km/sec, $V_s = 4.6$ km/sec). This model suggests that the crust under Tibet is thickened not by stacking two layers of a continental crust but by a complex process that might involve plastic deformation and vertical mixing.

The attenuation data complement the velocities for determining the crustal properties. Fitting the Rayleigh wave attenuation data of Bird and Toksöz⁽¹⁰⁾, part of which is shown in Fig. 3, requires an increased attenuation in the crust below about 45 km depth, although the thickness of this zone and the attenuation (Q_p^{-1}) value cannot be determined independently. The lowered velocities along with the increased attenuation suggest partial melting in the lower crust below the 45 km depth. Feng and He⁽¹⁶⁾ have also come to this conclusion for the eastern part of the plateau. Both the velocities and attenuation and the implied partial melting will serve as constraints for the thermal calculations in the next section.

EVOLUTION OF TIBET

How Tibet formed and how it evolved to its present state is an important problem. We will study this theoretically using finite difference calculations. Our models will start from the time of subduction of the oceanic plate on the leading edge of the Indian plate. Then it will go through the collision of India with Eurasia. After this, consequences of compression and crustal thickening in Tibet will be modeled in detail.

The possible sequence of events in the collision of the Indian and Eurasian plates is shown schematically in Fig. 9^(4,17,18,19).

The different stages of this process are:

1. The oceanic slab subducts under a continental lithosphere which may be "proto Tibet".
2. Subduction induces convection in the asthenosphere and heats the lithosphere from underneath.
3. The collision of India and Asia occurs and the Indian suture develops. Once the collision occurs, continental crust of India cannot be subducted because of its low density. The continued motion of India induces shortening of Tibet with the heated lower crust and lithosphere deforming plastically.
4. As Tibet rises the stresses required to lift it further become greater. After a critical elevation is reached, thrusting occurs on the colder Indian side of the plate. At this stage the Main Central Thrust (MCT) develops.
5. Subduction under the MCT continues until buoyancy forces become large. Then the Main Boundary Thrust forms.

6. During these times the convergence zone is still moving northward; Tibetan crust is thickening and becoming hotter at depth. The top half of the crust, however, remains elastic and transmits stress to Asia.

The regional thermal model based on this sequence of events was carried out using finite difference calculations with both convection and conduction included. The regional temperature profiles are shown in Fig. 7. The model shows that broken oceanic slab is still cool. Temperatures under the Himalayas are normal or below normal except in local regions of high strain heating. Lithospheric temperatures are higher than normal under Tibet. The geologic history is certainly much more complex and there are indications of different dips of subduction in different parts of the plateau⁽²⁰⁾. A southward dip of subduction would somewhat alter the mantle calculations, but the crustal calculations include a range of possible mantle temperatures wide enough to encompass this possibility. The detailed calculation for Tibet focuses on the crust.

The thickening of Tibet is modeled as being due to strain as the crust is thickened by internal deformation. The thermal history of the crust is modeled using a finite difference solution of the equation of heat conduction which includes the option of having material moving up and down due to internal deformation. To do this the crustal grid point spacing is increased while thickening is going on. The strain heating is turned off at the end of thickening.

Strain heating for an area being thickened is difficult to estimate and almost certainly involves both shear strain heating along faults and heating due to ductile flow. We treat only the case of heating along faults. This overestimates the heating if ductile flow occurs in the lower crust. The amount of heat produced in thickening is calculated for thrust faults dipping at 45°. The faults are taken to be distributed uniformly through the crust. Each grid point in the model gets the same amount of strain heat during the period of thickening. Two cases are considered in the modeling: $\tau = 1$ kb and $\tau = 2$ kb (τ = shear stress). In both cases the average heat production rate is less than that due to the radioactive heat sources.

The amount and distribution of radioactive elements in the crust is another important input to the models. Again two cases are tested. One has two layers in the crust of equal thickness, but with three times the abundance of heat producing elements in the upper crust, which is considered to be the normal case. The other case is for homogeneous distribution of radioactive elements. This homogenization could occur during the thickening by the mixing of the layers. In both cases the same average crustal heat production rate is assumed. It is taken to be 7.1×10^{-6} ergs/cm³.

In the modeling both a normal mantle temperature gradient and higher mantle temperatures are considered. The scheme includes 30 km of mantle lithosphere which is pushed down and heated by the crust. No heat sources are put in the mantle material, but at the base of the mantle layer a fixed temperature boundary condition is imposed to simulate the effect of convection. The boundary condition at the surface is that it is maintained at 0°C. The initial condition for all cases is the steady state temperature solution for the particular crustal arrangement tested.

Four examples of these models are shown in Fig. 8 which gives the depth to the isotherms shown against time during and after the thickening of the crust. Partial melting is taken to occur when lower crustal temperatures are above 700°C. We see that with uniform crustal radioactive abundances (1 layer crust), (Fig. 8a,b,d), low shear stress and normal mantle heat flux, we do get partial melting in the lower crust. We then see that concentrating the radioactive

heat sources in the upper half of the crust (Fig. 8c), even with doubled strain heating, leads to significantly lower temperatures. Next we see that the time span of the thickening has little effect on the final temperature profile (as illustrated in Fig. 8a). Of course, if most of the thickening took place in the last part of the 40 million year period since the collision, the crustal temperatures would be lower. Finally, the effect of a hotter mantle, giving slightly less than twice the mantle heat flux, greatly increases the lower crustal temperatures.

CONCLUSIONS

Available refraction data and the Rayleigh wave phase and group velocities independently and jointly require about a 70 km thick crust under Tibet. Average compressional and shear velocities in the crust are low and there is relatively little change of velocities with depth. The above properties suggest that the crust is thickened by some mechanism that may have involved vertical mixing. The high attenuation in the lower crust implies high temperatures and possibly partial melting below about 45 km depth.

The thermal consequences of the crustal thickening due to internal deformations, calculated using a finite difference scheme, indicate that redistribution of crustal radioactivity contributes most significantly to the heating of the thickened crust. Heat flux from the mantle and shear strain heating during crustal thickening are other factors that affect the temperatures, although the effect of shear strain heating is less important. In about 40 m.y., temperatures in the lower crust may rise by about 200°C over typical lower crustal temperatures. This could produce partial melting in the lower crust, and can explain the low velocities and high attenuation of seismic waves.

In the middle and upper part of the crust, the temperatures are about the same or less than would be seen in normal crust. To explain the regions of high heat flow which are implied by hot springs, recent volcanics and magnetic data, upward material and heat transport is required. The partial melting of the lower crust may allow hot material to move as diapirs into the upper regions of the crust, thus heating these areas. This may also contribute to the general manner in which radioactive elements are concentrated upward in the crust.

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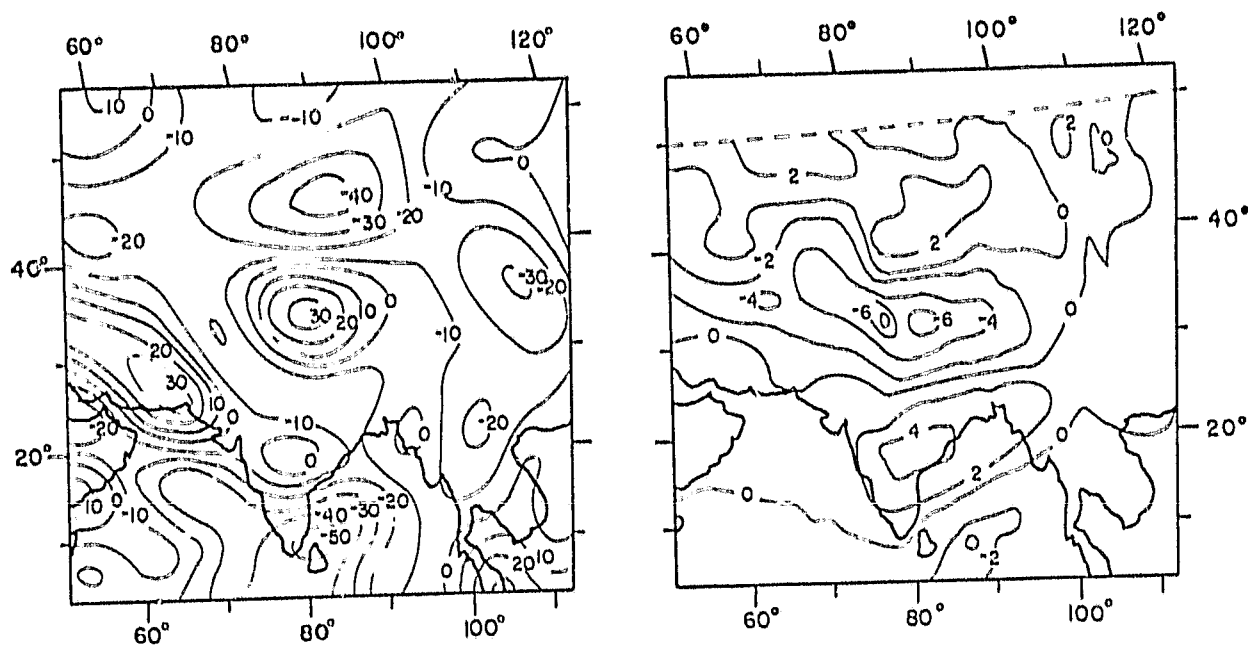


Figure 1. Comparison of gravity and magnetic anomalies over Tibet as measured by satellites. The heavy lines outline the landmasses and the latitudes and longitudes are also shown. On the left the free air gravity anomalies in milligals taken from the NASA Goddard Earth Model-8 are shown. The magnetic anomalies in gammas, for 500 km altitude(21), are shown at the right. Magnetic data above 50° north latitude are not reliable.

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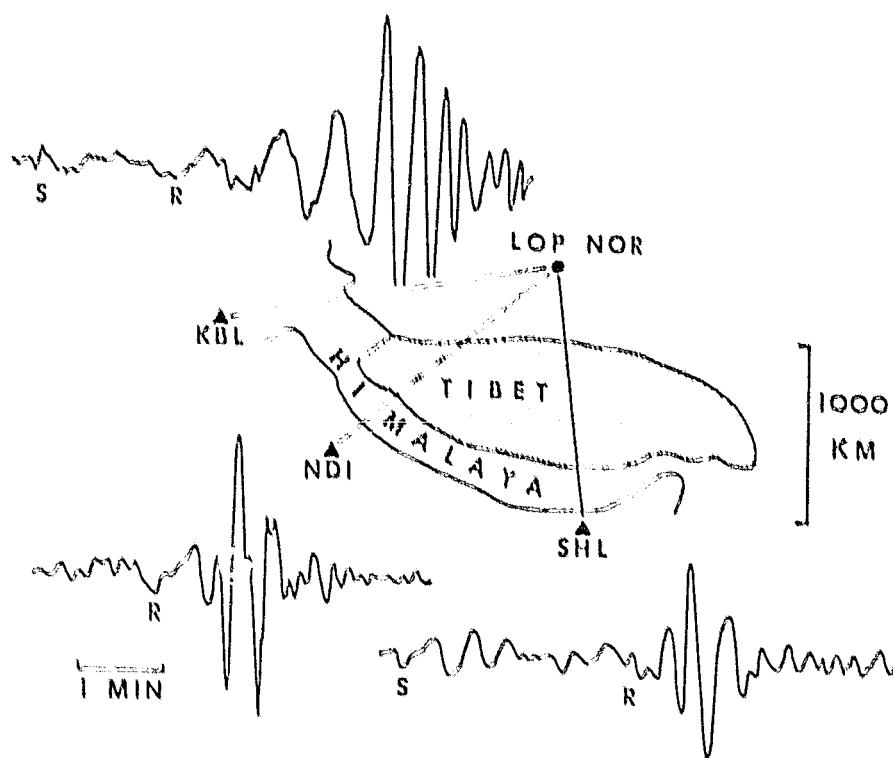


Figure 2. Rayleigh wave seismograms from an explosion at Lop Nor. Note the greater attenuation of the long period (10-50 second) waves which cross Tibet.

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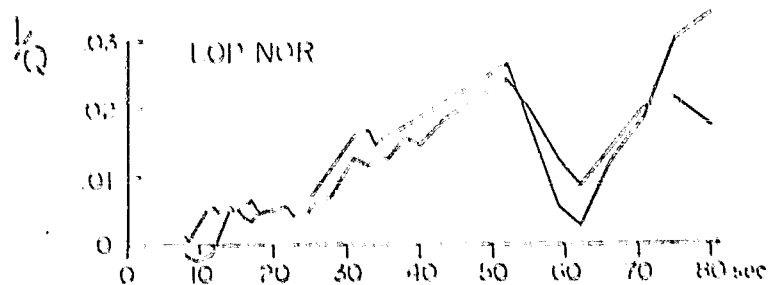


Figure 3. Observed attenuation ($1/Q$) of Rayleigh waves crossing Tibet for the Lop Nor explosion.

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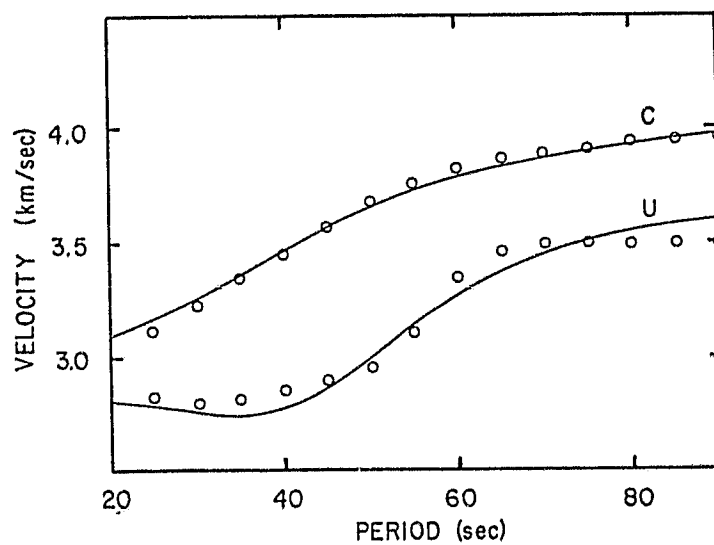


Figure 4. Rayleigh wave phase velocity (c) and group velocity (u) for the average of all paths crossing Tibet.

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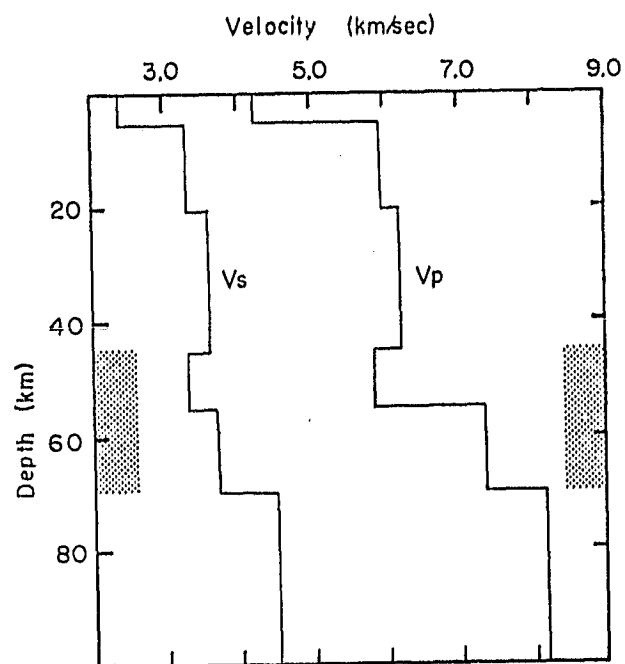


Figure 5. Crustal S-wave velocity model resulting from the inversion of Rayleigh wave data combined with the refraction P-wave profile (15) shown at right. Dotted area indicates the region of high attenuation.

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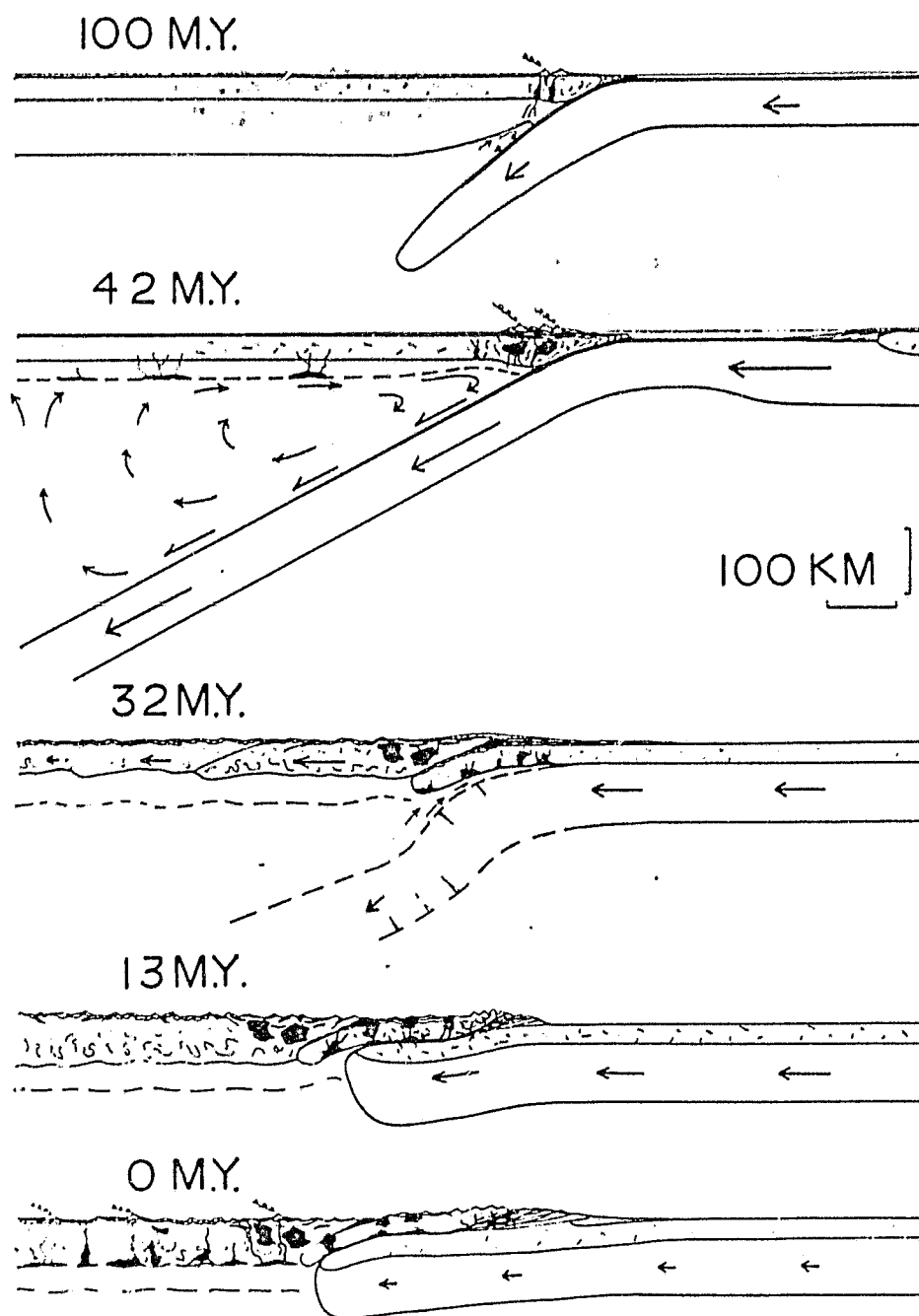


Figure 6. Schematic representing the evolution of the Himalayas and Tibet from the Cretaceous to the present.

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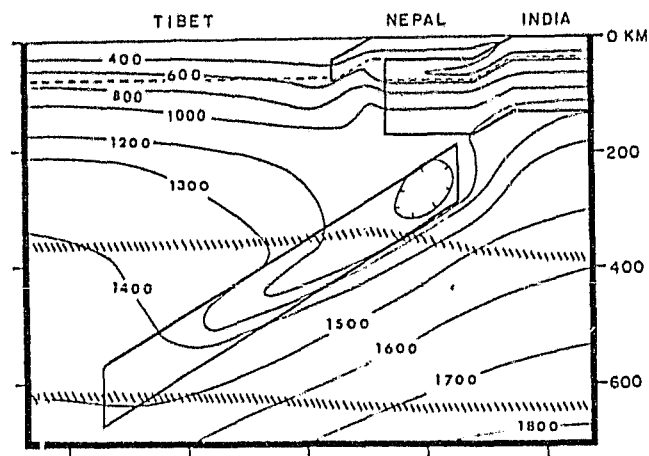


Figure 7. Present day temperatures in the Himalayan-Tibetan region that would result from the sequence of events shown in Fig. 6. The Moho is shown as a dashed line and upper mantle phase boundaries are shown as diagonal slashes. The broken slab is bounded by solid lines and is sinking. Note the general depression of temperatures under the Himalayas as a result of subduction of cold material. Temperatures are above normal under Tibet. From Toksöz and Bird(19).

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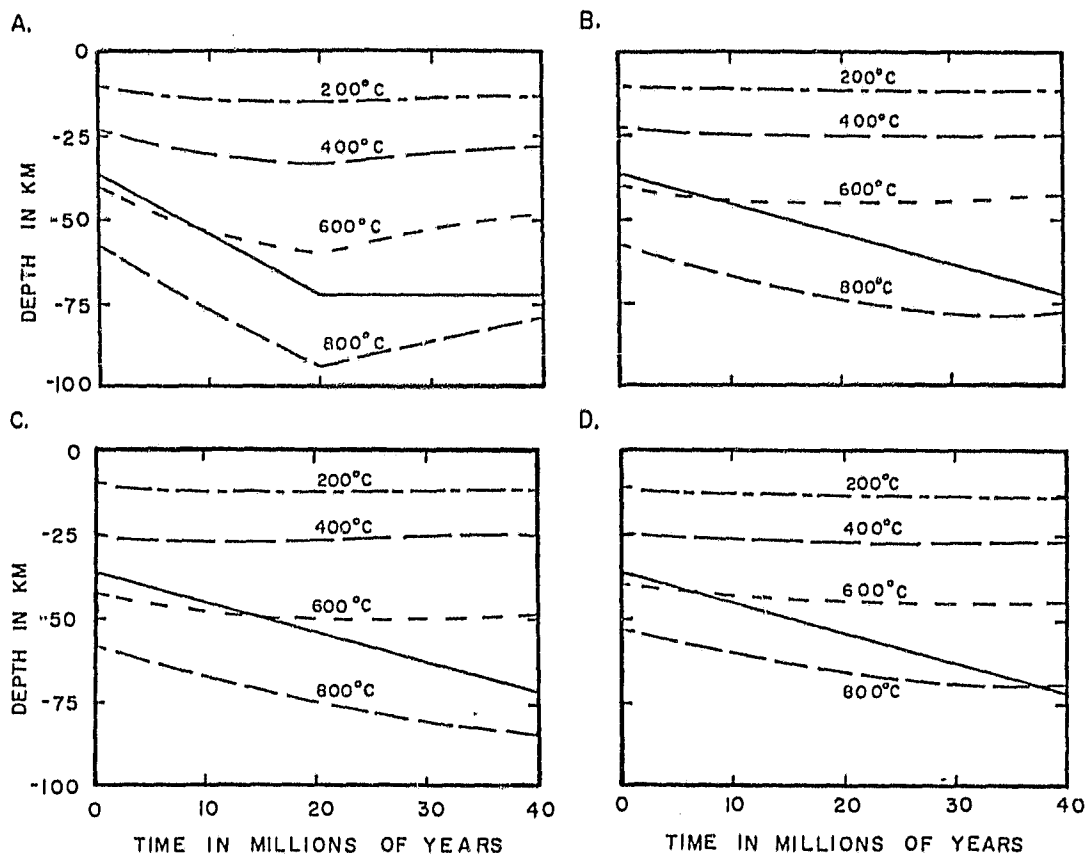


Figure 8. Results of the crustal thermal calculations. Dashed lines are depth to isotherms against time since the collision of India and Asia. Solid line is the crust-mantle boundary. All cases have the doubling in crustal thickness occurring in 40 million years except case A which has it taking 20 million years. Case A is for a 1 layer crust and 1 kb shear stress. Case B is for a 1 layer crust and 2 kb shear stress. Case C shows that a 2 layer crust even with 2 kb shear stress leads to much lower temperatures. Case D has one third more mantle heat flux and thus higher temperatures in the lower crust.

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III.2 CONVECTIVE INSTABILITY OF A THICKENED BOUNDARY LAYER AND
ITS RELEVANCE FOR THE THERMAL EVOLUTION OF
CONTINENTAL CONVERGENT BELTS *

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ABSTRACT

When crust thickens during crustal shortening, the underlying mantle lithosphere must shorten and thicken also, causing the submersion of cold, dense material into the surrounding aesthenosphere. For a range of physical parameters, the thickened boundary layer that forms the transition from the strong lithosphere to the convecting aesthenosphere may become unstable, detach, and sink into the aesthenosphere, to be replaced by hotter aesthenospheric material. We have studied the instability of a thickened boundary layer for a range of physical parameters (Rayleigh number), amounts of thickening, and boundary conditions. In all cases, the fluid was overlain by a rigid, conducting layer. Extensive numerical experiments were made for fluids with stress free boundary conditions, either heated from below or from within. From a simple physical description of the observed pattern of flow, we derived expressions that related the growth of the

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instability and the time needed to remove the thickened boundary layer as a function of the amount of shortening (f), the Rayleigh number (R), and the ratio (a/d) of the thicknesses of the rigid and fluid layers. In our opinion, observations and theory agree well (within 10% for $R > 10^5$) and show that the speed with which the thickened boundary layer is removed increases with increasing f , R and a/d . A limited series of runs with no slip boundary conditions suggests approximately the same functional relationships, but with the process 0 to 30% slower than with stress free boundaries. For Rayleigh numbers comparable to those appropriate for upper mantle convection (10^5 - 10^7) the removal of the boundary layer occurs rapidly, in times less than the thermal time constant of the overlying rigid plate. Using typical values for the physical parameters in the earth, the boundary layer is removed in times less than the duration of deformation in some collision zones (30-50 my). Thus, we suspect that often the lower lithosphere is removed during the process of crustal shortening, causing the overlying crust and uppermost mantle to warm rapidly. This process is likely to contribute to the development of regional metamorphism and to the generation of late-or post-tectonic granites. We suspect, in fact, that in some cases the entire mantle lithosphere may detach from the lower crust during crustal shortening, exposing the crust to aesthenospheric temperatures.

III.3 CONVECTIVE INSTABILITIES IN A VARIABLE VISCOSITY
FLUID COOLED FROM ABOVE *

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ABSTRACT

We study the influence of the viscosity structure on the development of convective instabilities in an infinite Prandtl number fluid cooled from above. A viscosity (ν) dependence with depth (z) of the form $\nu_0 + \nu \exp(-\gamma z)$ was assumed. After the temperature of the top boundary is lowered, velocity and temperature perturbations are followed numerically until convective breakdown occurs. Viscosity contrasts of up to 10^7 and Rayleigh numbers of up to 10^8 were studied.

For intermediate viscosity contrasts (around 10^3), convective breakdown is characterized by the almost simultaneous appearance of two modes of instability. One involves the whole fluid layer and has a large wavelength. The other mode has a much smaller critical wavelength and develops below an upper mechanical boundary layer which behaves rigidly. The "whole layer" mode dominates for small viscosity contrasts but is suppressed by viscous dissipation at large viscosity contrasts.

For the "rigid top" mode of instability, we propose a simple method to define the rigid lid thickness. We are thus able to compute the true depth extent and the effective driving temperature difference of convective flow. A measure of viscosity is provided by the dissipation-weighted average of viscosity throughout the fluid. Because viscosity contrasts in the convecting region

rarely exceed 10, simple scaling arguments are sufficient to describe the instability. The critical wavenumber scales with the thermal boundary layer thickness beneath the rigid lid. Convection occurs when a Rayleigh number defined locally (using this boundary layer thickness as length-scale) exceeds a critical value of 160-190. The critical value depends on the thickness of the rigid lid.

The local Rayleigh number may be computed at any depth in the fluid. Convection develops below depth z_R (the rigid lid thickness) such that this local number is maximum in the layer.

If the age of 70 My which marks the flattening of the depth vs. age curve on the ocean floor is the critical time for the onset of convection beneath the plate, the above results can be used to constrain the average value of viscosity in the thermal boundary layer to be around $3 \times 10^{16} \text{ m}^2/\text{s}$.

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III.4 On Melting of the Subducted Oceanic Crust Beneath Island Arcs

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ABSTRACT

The source of island arc magma has received extensive attention. Based on petrological and geochemical arguments, it is possible that arc magma is derived from subducted oceanic crust. Previous thermal models, however, seem to suggest that it is very difficult, if not impossible, to melt a cold, subducted oceanic crust at sufficiently low pressure where island arc magma is believed to be derived. This misconception is created because plate subduction is generally interpreted as a cooling process. This interpretation is supported by most of previous thermal models which are calculated without considering the dynamic effects of the wedge of mantle between the subducting and arc plates. A consideration of the dynamics of the mantle above the plate, suggests that subduction will generate an induced flow in the wedge above the subducting slab. This current continuously feeds hot mantle material into the corner and onto the slab surface. A high temperature thermal environment can be maintained in the vicinity of the corner, immediately beneath the over-riding plate. Our regional models further demonstrate quantitatively that the existence of an induced flow is probably able to produce local melting of a subducted oceanic crust just about 30 km down dip from the wedge corner. Additional geological processes such as reasonable amounts of shear heating and minor dehydration will further increase the probability of melting a cold subducted oceanic crust at shallow depth.

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IV. SEISMICITY, STRESS AND DEFORMATION MODELS OF CONTINENTAL CONVERGENCE ZONES

The convergence and the collision of continental plates produce large stresses in the lithosphere. In the crust, these stresses produce elastic and creep type deformations and brittle fractures manifested as earthquakes. Mechanisms of large earthquakes provide a good definition of the stress regime and relative motions of the plates at the boundaries. In addition to horizontal forces, under the Himalayas the underthrusting Indian Plate is subject to large vertical forces due to the weight of the mountains. The effect of this load on the flexure of the plate could be significant.

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IV.1 MECHANICAL RESPONSE OF LITHOSPHERE DUE TO UNDERTHRUSTING AT THE HIMALAYA

Hélène Lyon-Caen

To constrain the mechanics of continental collision, we are studying the flexure of the Indian shield due to the load of the Himalaya. Seismic and gravity data suggest that as India underthrusts the Himalaya, the crust thickens northward beneath the range. We are analyzing the gravity anomalies there, assuming that, first, the shape of the Moho is simply the response of a flexed plate (India) to the weight of the mountains on it and, second, that the gravity anomalies are due primarily to the density differences among crust, mantle and sediments in the Ganga Basin. Assuming a density for the sediments in the Ganga Basin, south of the Himalaya, and a density contrast between the crust and mantle, and the load of the mountains, we calculate the position of the top surface of the Indian plate and a corresponding free air gravity anomaly profile. We can vary the flexural rigidity of the plate, the position of the northern end of the plate, the density contrasts among sediments, crust, and mantle, and the bending moment applied to it. Preliminary calculations show the addition of a bending moment to the India plate 100 km to the north of the Himalayan front has a negligible effect on the shape of the Moho and the calculated gravity anomalies. We are in the process of systematically examining the effects of the density differences, flexural rigidity of the plate, and position of the northern

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end of the plate. The figures show the gravity profiles obtained for various models.

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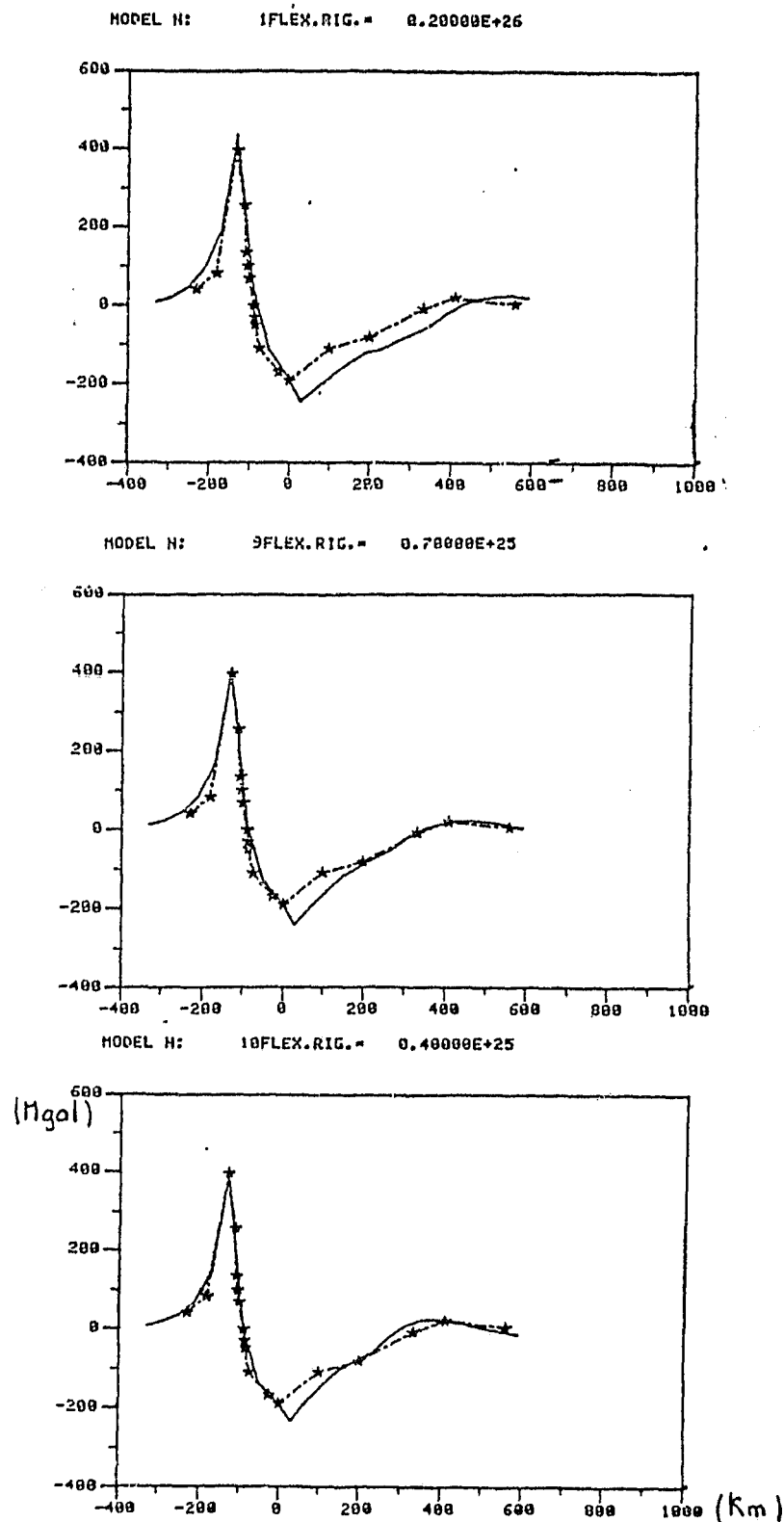
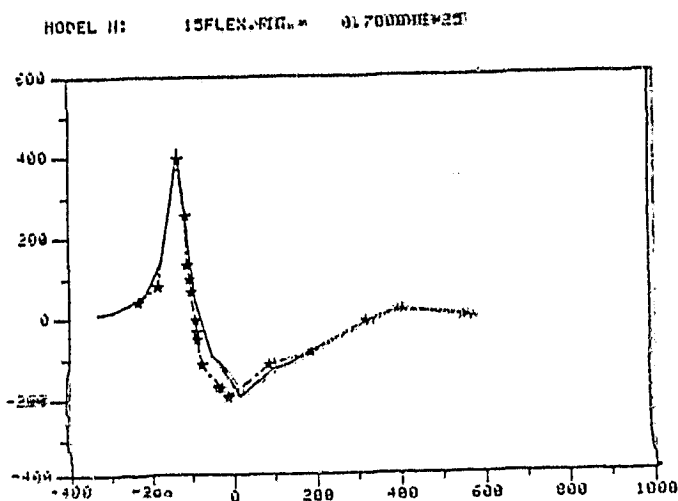
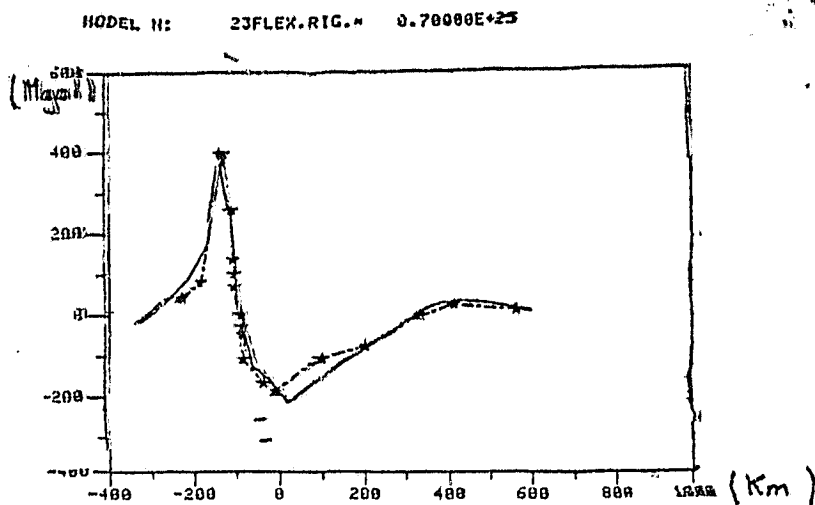


Figure 1: Free air anomalies profiles calculated for different values of the flexural rigidity D (continuous line). The stars (*) represent the data. From the top, $D = 210^{25}$, $.710^{25}$, $.410^{25}$ N.m. The point $x=0$ is at the Himalayan front. The density contrast crust-mantle, $\Delta\rho$, is $.55 \text{ g/cm}^3$ and the northern end of the plate is at $x_g = -200 \text{ km}$.

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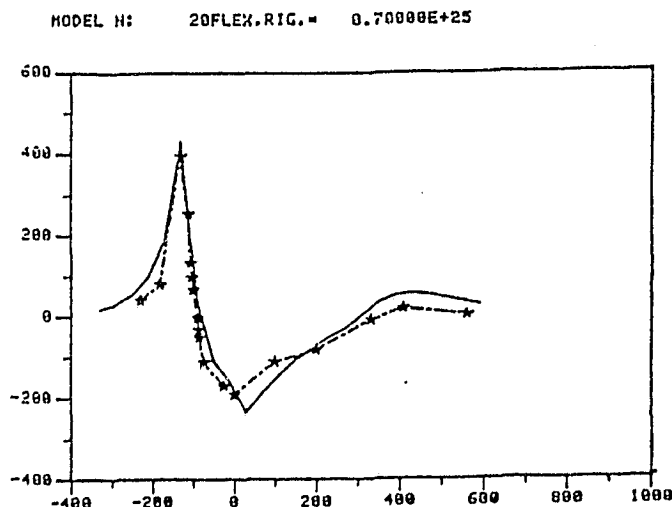


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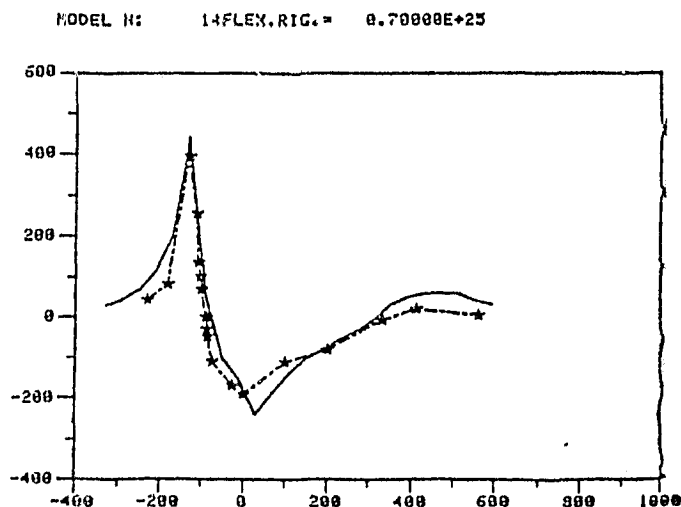


2-b

Figure 2: Free air anomalies profile calculated for different position x_E of the northern end of the plate: $x_E = -150$ km for 2.a; $x_E = -600$ km for 2.b. $D = .710^{25}$ N.m. and $\Delta\rho = .55$ g/cm³ in the two cases.



3-a



3-b

Figure 3: Free air anomalies profile calculated for different values of $\Delta\rho$: $\Delta\rho = .45$ g/cm³ in 3.a; $\Delta\rho = .6$ g/cm³ in 3.b. $D = .710^{25}$ N.m. and $x_E = -200$ km.

$\Delta\rho = .55$ g/cm³ and the northern end of the plate is at $x_E = -200$ km.

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IV.2 MECHANISMS OF SOME LARGE EARTHQUAKES IN THE ALPINE-HIMALAYAN CONVERGENCE BELT

M. Nafi Toksöz and John Nabelek

Large earthquakes provide important information about the present tectonic deformation and the associated state of stress in the lithosphere. The larger the earthquake size, the more likely it represents the average tectonic deformation in a given area. In order to assess the present state of stress, the determination of the earthquake magnitude, moment, fault dimensions and slip vector is essential. In this regard, analysis of the teleseismic surface waves and waveform modeling of body waves, combined with in situ geological observations, have proven to be quite effective. Our study included three large events, the Lice (1975) earthquake in southeastern Turkey, the Caldiran (1976) earthquake on the Turkish-Iranian border, and the recent El-Asnam (1980) earthquake in Algeria.

The Lice earthquake occurred south of the East Anatolian Fault and had essentially a thrust source mechanism with some left-lateral component. This is consistent with the northward motion of the Arabian plate. The Caldiran earthquake occurred east of the junction between the North and East Anatolian Faults, on a previously unrecognized fault. Surprisingly, it had a pure strike-slip mechanism, in an area where thrusting was thought to be the major mode of deformation. The El Asnam earthquake is consistent with the northward motion of Africa relative to Europe.

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Table 1 summarizes the source parameters of these three events. Surface breaks, source of the motion and fault plane solutions of these earthquakes are shown in Figures 1-3.

TABLE 1

Source Parameters of Recent Large Earthquakes in the Alpine-Himalayan Belt

Earthquake and Date	Epicerter Latitude Longitude	Fault Length (km)	Magnitude M _S	Moment dyne-cm	Ave. Slip m	Stress Drop (bars)	Strike	Dip	Slip Dir.
Lice Turkey 9/6/75	38.47 40.72	20	6.7	9x10 ²⁵	0.5	15	255°	52°N	36°
Caldiran Turkey 11/24/76	39.12 44.03	55	7.3	7x10 ²⁶	2	35	290°	78°S	176°
El Asnam Algeria 10/10/80	36.13 1.40	40	7.3	5x10 ²⁶	4.4	97	217°	52°N	82°

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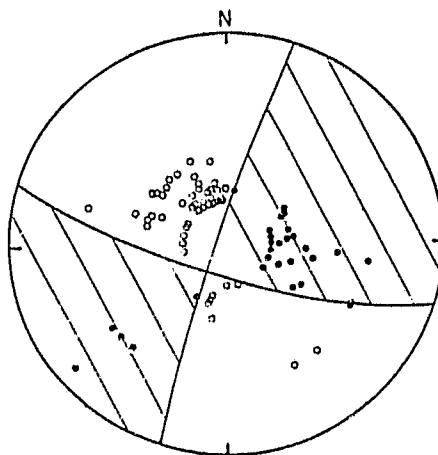
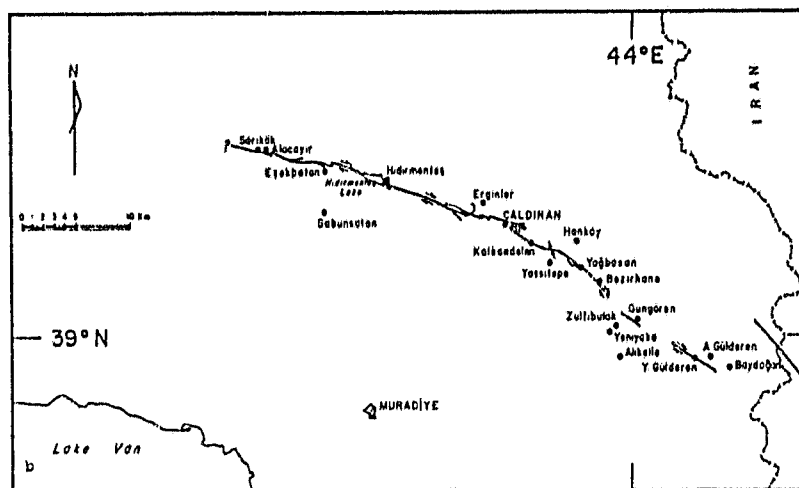


Fig. 2 - The observed fault trace of the Caldiran (1976) earthquake and the corresponding fault plane solution.

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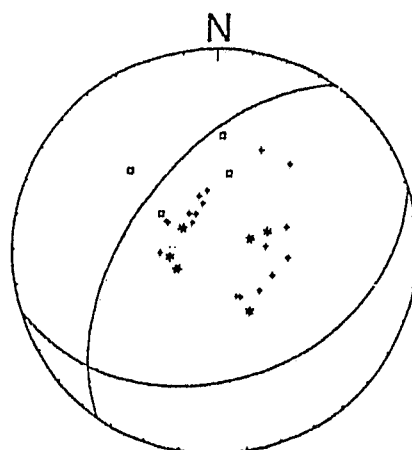
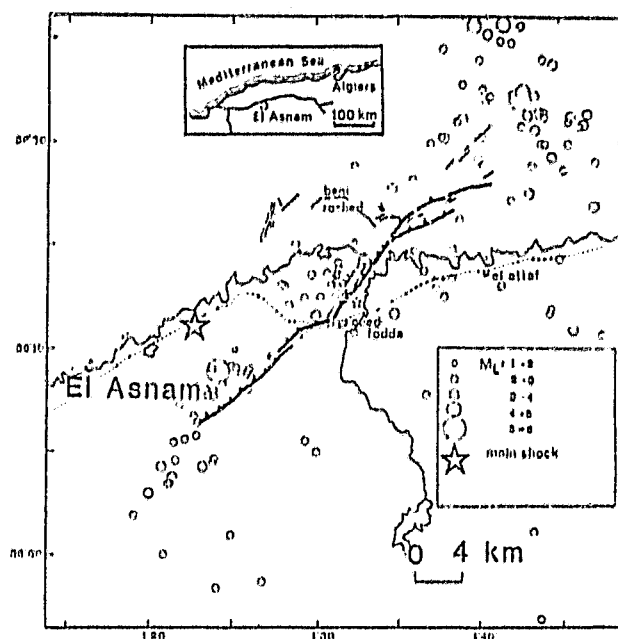


Fig. 3 - The observed fault trace of the El Asnam (1980) earthquake and the corresponding fault plane solution.

IV.3 SEISMICITY AND MOUNTAIN BUILDING *

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of Technology, Cambridge, Massachusetts 02139)

We summarize some basic aspects of the seismicity of several individual mountain belts and of much of eastern Asia, with the point of view that seismicity places an important constraint on the physical processes occurring during mountain building. Much of the seismicity of the Himalaya seems to indicate that at present the India plate slides beneath the Himalaya as a coherent entity along a shallow plane. Although some seismicity beneath the Himalaya suggests internal deformation of the overlying mountains or of the region beyond the zone where the India plate remains coherent, at present there does not seem to be active detachment of a crystalline nappe from India, as seems to have occurred earlier at the Main Central and Main Boundary faults. In the Zagros, the seismicity indicates a shortening of the basement beneath the overlying, probably detached, sedimentary cover. If Arabia is sliding beneath Iran along a shallow dipping fault, it does so aseismically. The Tien Shan experiences north-south shortening by thrust faulting and strike-slip faulting caused by the convergence of the Tarim basin to the south and the Paleozoic platform to the north. Seismicity and deformation occur throughout the belt, with the stable areas north and south of it underthrusting the mountain belt. Seismicity in the Peruvian Andes concentrates on the flanks of the mountains. As in the Tien Shan, the belt is two-sided and experiences thrust faulting on both the east and west flanks. Moreover, in both areas, fault planes are steep, suggesting a clear analogy with the Colorado and Wyoming Rocky Mountain in the early Tertiary than with either the Himalaya at present or the Canadian Rockies and the Sevier belt in the western U.S. in the late Cretaceous. Active normal faulting at high altitudes in the Andes, however, makes them differ from the Tien Shan. The normal faulting is probably a result of buoyancy forces acting on the elevated areas. The northwest Himalaya, Pamir and Hindu Kush in Pakistan, Tadjikistan and Afghanistan show yet more varied and more complicated deformation than the belts described above. A spectrum of types of faulting and widespread seismicity and deformation characterize the region. In many ways it seems to be a miniature

The seismicity of eastern Asia indicates active deformation in a region 1000 to 3000 km wide between the more stable, essentially aseismic Indian and Eurasian plates. Some of this seismicity and associated deformation is simply and directly attributable to the convergence of these plates. The thrusting in the Himalaya is clearly a result of this convergence, and that in the Tien Shan can also be interpreted as a consequence of it. Seismicity north and east of Tibet indicated a spectrum of fault types that include strike-slip and even normal faulting. Nevertheless the relatively consistent orientations of P and T axes over large areas are consistent with a stress field generated by India's penetration into Eurasia, and accordingly the seismicity can be considered a direct consequence of this penetration. The scattered earthquakes in Tibet, as well as the Quaternary faulting, indicate active normal faulting with east-west extension. This style of deformation probably results from buoyancy forces acting on the elevated and thickened crust of Tibet. Thus the active tectonics of Tibet are due to the penetration of India into Eurasia that caused the high elevations and thick crust, but this crustal shortening seems to have stopped except on the flanks of Tibet.

A study of depths of foci of earthquakes shows that most events occur in the cold outer 15 km of the crust. Deeper events seem to occur in older, colder, more stable shields, in belts where thrust faulting and crustal thickening have advected cold material down, or in the underlying mantle. The lower crust, however, seems usually to be aseismic and probably deforms ductilely. We infer that it is weaker than the region above and below it.

The seismicity of mountain belts is clearly more widespread and deformation is more varied than in oceanic regions. Several simple and obvious factors contribute to the greater complexity of continental regions and to the difficulty in gaining a quantitative description analogous to that of plate tectonics in oceanic regions. One surely is the inadequacy of the historic record of seismicity in portraying the long term seismicity. A second is the large lateral variations in strength due both to lateral variations in temperature and rheology and to inherited zones of weakness. A third is the effect of different strengths of the minerals abundant in the crust and mantle, which may introduce a zone of low strength between the middle crust and uppermost mantle in continental but not oceanic regions. Finally, the stresses that balance the gravitational body forces must vary from regions of high to low elevation and of

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thin to thick crust, thus perturbing the regional stress field. While it is easy to recognize these sources of complexity, our problem for the future is to obtain data that constrain them quantitatively.

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INTRODUCTION

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In keeping with the basic geologic doctrine that "the present is the key to the past", our understanding of mountain building is likely to grow rapidly with the study of active belts. At the same time a thorough understanding requires a knowledge of the deep structure and of the processes that occur there. Although much of this information is provided by detailed studies of exhumed belts, we view seismicity as a link between geologic studies of active faulting and those of deeper structure by providing information about the active processes occurring beneath the earth's surface.

For us the word seismicity includes all geologic aspects of earthquakes --not only where and how frequently they occur but also the orientation, extent and amount of faulting involved in particular earthquakes. In Appendix A we give a general discussion of uncertainties and pitfalls in the interpretation of seismic data. As on occasion there has been misuse of seismic data, we think that it is useful to give the reader some simple rules for the cautious use of published seismic data.

This paper is concerned only with those mountain belts that are due directly or indirectly to large-scale convergence of continental masses. We do not discuss the seismicity of the continental rift zones, such as in east Africa, which although mountainous are not related to convergent plate margins. Similarly, our attention is not directed towards those subduction zones where one slab of oceanic lithosphere

underthrusts another. There are two main differences between the seismicity of subduction zones and that of continental convergent zones. At subduction zones, usually there is a thin, shallow dipping ($\sim 10^\circ$ - 30°) planar zone of shallow seismicity ($h < 70$ km) that defines a major thrust fault along which underthrusting takes place. Then at greater depths the planar seismic zone gradually steepens, usually to about 45° , but sometimes to 90° . These intermediate and deep focus earthquakes, however, occur within the subducted oceanic lithosphere, not on its upper boundary (e.g., Isacks and Molnar, 1971). Generally, seismicity is more diffuse in continental regions and there is no zone of intermediate and deep focus earthquakes like those at island arcs. The shallow seismicity often includes a spectrum of fault plane solutions and shows a much more complicated pattern of deformation than the simple underthrusting at island arcs. The few inclined zones of intermediate and deep focus seismicity within continental areas (in the Pamir-Hindu Kush, Burma, and the Carpathians) are usually cited as evidence for subduction of oceanic lithosphere in the last 10 to 20 my. Recently, intermediate depth seismicity has been recognized in other active convergent zones, but this activity seems to be part of the diffuse deformation of continental convergent zones at depth and does not reflect subduction of oceanic lithosphere (Chen et al., 1981; Chen and Molnar, 1981a, b; Chen and Roecker, 1980; Hatzfeld and Frogneux, 1981).

Below, we first discuss the seismicity of several active mountain ranges, beginning with the Himalaya, which seems to us to be one of the simplest. Then we consider broad convergent zones and we give a brief discussion of possible variations in seismicity during the development of mountain belts. Finally, before giving a brief summary, we discuss the relevance of depth of earthquakes to the crustal and mantle rheology.

LINEAR MOUNTAIN RANGES AND CONVERGENT PLATE BOUNDARIES

Often plate boundaries in continental regions are marked by long, narrow mountain ranges. These ranges (and boundaries) do not necessarily separate two essentially rigid plates, but instead they often mark the boundary of one plate with a broader zone of deformation that separates that plate from its nearest neighbor. Two clear active examples of such belts are the Himalaya and Zagros, which bound the Indian and Arabian shields on the north and northeast. These belts are relatively narrow and the style of deformation changes little along strike. The Tien Shan and the Peruvian Andes are somewhat broader zones of deformation but separate two essentially rigid converging blocks or plates. Finally the northwestern Himalaya, the Hindu Kush, and the Pamir in Pakistan, Afghanistan and Tadjikistan comprise a yet broader zone with little evidence of linearity. We discuss each of these individually and make comparisons as each new belt is discussed.

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The Himalaya

Among the mountain belts discussed here, the Himalaya, from its northeast corner of Assam to Kashmir, is probably the simplest. Seismicity located teleseismically follows the range and most of it lies between the Main Central Thrust and the Main Boundary Fault (Figure 1) (see Gansser, 1964, 1977; or LeFort, 1975, for geologic background). Most fault plane solutions show underthrusting either along shallow planes dipping north beneath the mountains or along steep planes dipping south (Figures 1 and 2) (e.g., Armbruster et al., 1981; Fitch, 1970; Molnar et al., 1973, 1977). In accordance with the dips of geologically mapped thrust faults in the Himalaya, the northerly dipping planes are probably the fault planes, and the earthquakes reflect northward convergence between the Indian shield and the Himalaya. The shallow dipping nodal planes dip north at angles of 15° or less in eastern Nepal and Assam, but some dip more steeply ($\sim 30^\circ$) in western Nepal and further west (e.g., Armbruster et al., 1981; Molnar et al., 1977). Thus, in the east the solutions seem to include a coherent underthrusting of India beneath the Himalaya (Molnar et al., 1977). Some of those farther west seem to indicate internal deformation of the Himalayan crust (Armbruster et al., 1981; Seeber et al., 1981).

Unfortunately, we are not aware of any detailed studies using local networks of particular portions of the Himalaya between Kashmir and Assam. Thus, the depth distribution of earthquakes is poorly known. A study of some of the larger

events, however, shows that they occur at shallow depths, 15-20 km (Armbruster et al., 1981). These depths and the shallow dipping planes are consistent with the events, at least in the eastern part of the range, occurring along the interface between the underthrusting Indian shield and the overriding mountain mass (Figure 3).

The historic seismicity of the Himalaya includes four major events ($M \geq 8$) in the last 100 years (Figure 1). The fault associated with one, the 1897 Assam earthquake (Oldham, 1899; Richter, 1958), however, probably underlies the Shillong plateau but not the Himalaya. Moreover, controversy surrounds the 1950 Assam earthquake. Whereas we suspect the important displacement occurred along a shallow, north dipping plane (Chen and Molnar, 1977), Ben-Menahem et al. (1974) inferred slip on a north-northwest striking plane. Unfortunately, little is known about the slip associated with either the 1905 Kangra earthquake or the 1934 Bihar-Nepal earthquake. From the fault plane solutions of nearby events, and from the intensity distributions associated with them, it is likely that those two events resulted from low angle underthrusting of India beneath the Himalaya (Armbruster et al., 1981; Molnar et al., 1977; Seeber et al., 1981). The occurrence of four major events in the last 100 years attests to important continuing tectonic activity in the region, but too little is known about any of these events to alter or verify the simple interpretation deduced from the focal depths and fault plane solutions discussed above.

The tectonic setting at the Himalaya is very similar to that of an island arc (e.g., Molnar et al., 1977). The Indian plate bends down in front of the Himalaya and forms a deep basin, the Ganga basin. One event beneath it, in fact, shows normal faulting with a T axis perpendicular to the range (Molnar et al., 1973, 1977; Figure 1). This event is presumably analogous to those beneath deep sea trenches (e.g., Chapple and Forsyth, 1979; Stauder, 1968) and therefore is probably due to flexing of the Indian plate as it bends down in front of the Himalaya. The flexed Indian plate supports some of the excess mass in the Himalaya and maintains parallel belts of negative and positive isostatic anomalies over the Ganga basin and the Himalaya (H. Lyon-Caen, in preparation; Molnar et al., 1977; Warsi and Molnar, 1977). The Indian plate underthrusts northwards beneath the Himalaya along a fault or faults, dipping gently toward the north (Figures 2 and 3). At present, movement seems to us to be on the Main Boundary Thrust (MBT) or nearby parallel faults (Figure 3).

The geologic history of the Himalaya suggests that at least two large slivers of India were scraped off of its northern margin to make the Himalaya. One lies above the Main Central Thrust and the other between the MCT and the MBT (Figure 3). The depths of the earthquakes and fault plane solutions suggest to us that at present India is sliding intact beneath the Himalaya and that detachment of another sliver of India's basement is not taking place (Armbruster

et al., 1981). Presumably this detachment of slivers (or crystalline nappes) occurs only occasionally in the evolution of mountain belts, when sufficient underthrusting has taken place along the active thrust.

The Zagros

In many ways the tectonic settings of the Zagros and the Himalaya are similar. Both involve continental collisions beginning some time since the late Cretaceous. In both a shield has moved rapidly towards an Andean margin and has underthrust it towards the north or northeast. In neither are there well located intermediate depth events that could attest to recent subduction of oceanic lithosphere. However, whereas the collision between India and Eurasia seems to have occurred in the Eocene, the timing of that between Arabia and Eurasia remains more controversial with views ranging from late Cretaceous (e.g., Stöcklin, 1977; Takin, 1972; Tapponnier, 1977) to early Miocene (e.g., Dewey and Sengör, 1979). The seismicity does not clarify this, but the Zagros and the Himalaya do seem to deform differently.

Seismicity in the Zagros appears to be shallow, despite some assertions to the contrary. All careful studies of earthquake locations have failed to reveal activity deeper than about 25 km (Jackson, 1980a; Jackson and Fitch, 1981; Niazi et al., 1978). Fault plane solutions of larger events indicate relatively steep planes, dipping at 30° to 60° (Jackson, 1980b; McKenzie, 1972) (Figures 2 and 4). Thus

if there is a feature like the main boundary fault in the Zagros, it is apparently aseismic (or has been at least for the last century). Moreover, unlike the Himalaya, where there have been several major events ($M > 8$) during the last century, the largest events in the Zagros seem to be much smaller ($M < 8$). In fact, it is clear that much aseismic deformation must occur between Arabia and Eurasia. North (1974) summed the seismic moments for large events in Iran to obtain a rate of seismic slip for the last 70 years. He found that slip during earthquakes could account for only a few percent of the slip between Arabia and the stable parts of Eurasia implied by plate motion for the last few million years. North's (1974) result does not mean that the aseismic slip has occurred on a plane at a shallow dipping angle beneath the Zagros, instead of, for instance, on steeper faults north of the range, but perhaps such a possibility should not be abandoned yet.

Geologically one of the difficulties in studying the Zagros is that at the surface one sees folding and faulting of sediments that are probably detached from the basement along evaporite horizons. The depths of the earthquakes (Jackson and Fitch, 1981; Niazi et al., 1978) imply that most of the seismicity occurs in the basement below the thick overlying sediments. The areal distribution, depths and fault plane solutions of the earthquakes suggest that the whole belt experiences horizontal shortening of the basement and not a simple underthrusting of Arabia beneath Iran (Jackson and Fitch, 1981) (Figure 4). Jackson (1980b) in fact suggests

that this thrusting reactivated normal faults that formed in an earlier phase of extension. He notes that by such a mechanism hundreds of kilometers of crustal shortening could occur without creating unusually thick crust (Helwig, 1976). In any case, the seismicity and active deformation of the Zagros are quite different from those of the Himalaya.

The Tien Shan

The Tien Shan is an active belt between 200 and 300 km wide. Part of it separates two aseismic regions: the southern part of the Eurasian plate, a stable platform that was consolidated in the Paleozoic, and the Tarim basin, a Precambrian Shield. Active faulting within the Tien Shan is primarily thrust and strike-slip (e.g., Molnar and Tapponnier, 1975; Ni, 1978; Shirokova, 1967, 1974; Tapponnier and Molnar, 1979; Trifonov, 1978). Earthquakes and active deformation are not confined to the edges of the belt, but seem to occur throughout it (Figure 5). This includes three large earthquakes in the latter part of the last century and the early part of this one (e.g., Chen and Molnar, 1977). Only one of these events, the 1911 Kebin earthquake, is well studied (Bogdanovich et al., 1914). Clear reverse faulting of several meters was observed along segments of an east-west trending zone of surface deformation, 200 km long. This event, like the Chilik earthquake of 1887, occurred within the interior of the Tien Shan and not on its flanks. The topography of the belt shows numerous elevated blocks separated by basins elongated parallel to the range. The seismicity is not

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sufficiently well studied to show that these blocks are aseismic and move as rigid bodies with respect to one another or not, but some basins have shown little activity in the historic record (Simpson et al., 1981).

Most fault plane solutions show thrust faulting, apparently with P axes trending approximately north-south (Figure 5) (Ni, 1978; Shirokova, 1967, 1974; Tapponnier and Molnar, 1979). The solutions are not well constrained, but thrust faults mapped geologically or inferred from the topography seen on the Landsat imagery show strikes nearly parallel to the belt. The logical inference is that the crust is being shortened in a north-south or north-northwest - south-southeast direction by thrust and reverse faulting. Although the belt is not perfectly symmetrical, it has no clear polarity. The mountains seem to overthrust onto both Eurasia and Tarim as these aseismic blocks move towards one another. Dips of fault planes are generally steep (30° - 60°), although a few may be as low as 20° (Figures 2 and 5). Thus, in general, low angle thrusting as in the Himalaya, or decollement of thin sheets, does not seem to be occurring, at least on a large scale. Most depths of foci of larger events ($M \sim 5.5$ - 6.5) seem to be between 10 and 20 km (Romanovicz, 1981; Vilkas, in preparation). Nevertheless, the depth of at least one event is unusually deep, 40-50 km, and another occurred between 25 and 30 km (Chen and Molnar, 1977; Vilkas, in preparation). Therefore, the crustal basement, and possibly the uppermost mantle, are involved in this

deformation. Presumably the steep planes cut through the basement instead of flattening at greater depths. Thus, the style of deformation seems to resemble more closely that of the Zagros or that of the Rocky Mountains of the U.S. during the early Tertiary (Laramide) than that of either the Himalaya or the Canadian Rocky Mountains in the Late Cretaceous.

The active tectonics of the Tien Shan are complicated by the presence of major northwesterly trending right lateral strike-slip faults (e.g., Tapponnier and Molnar, 1979; Trifonov, 1978). These faults have clearly been active during the Quaternary (e.g., Burtman, 1963; Trifonov, 1978; Voitovich, 1969), but they probably are reactivated Paleozoic strike-slip faults (Burtman, 1961, 1963, 1975). We are not aware of any evidence for active seismicity along them (i.e., within the last 50 years) (see also Simpson et al., 1981), but the clear Quaternary displacements make it unwise to conclude that their activity has ceased. The orientation and sense of slip along these faults is consistent with a north-south or north-northwest -south-southeast maximum compressive stress, and with the idea that the Tien Shan is experiencing shortening with approximately the same orientation throughout the belt.

The Peruvian Andes

The close proximity of a subduction zone makes the Andes different from the other mountain ranges considered here. Yet, because of their tectonic similarity to intracontinental belts (e.g., Audebaud et al., 1973), discussion of them here seems appropriate .

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On the west side, active subduction of the Nazca plate creates most of the features typical of island arc structures. Fault plane solutions attest to eastward subduction (Stauder, 1975), and an inclined seismic zone dips at a shallow angle beneath the Andes (e.g., Barazangi and Isacks, 1976, 1979). The Andes, however, differ from most other subduction zones by the presence of a range of high mountains along the zone.

The origin of the Andes is controversial, but at least part of the high elevations and thick crust (see James, 1971; Ocola and Meyer, 1973) is surely due to east-west crustal shortening of the western margin of South America. Folding and thrust faulting in the Andes have dominated the structural evolution, at least since the late Cretaceous (e.g., Audebaud et al., 1973; Dalmayrac et al., 1981; Mégard, 1978). Mesozoic and Paleozoic sediments in the high Andes reflect this deformation and Pliocene and Quaternary sediments in the sub-Andes, on the eastern flanks of the range, are folded and overthrust one upon another.

The seismicity along the flanks of the Andes corroborates the suggestion of continued crustal shortening. Earthquakes seem to be shallow and most seem to occur in the crust between 10 and 25 km depth (Chinn and Isacks, 1981; Suárez et al., 1981). Two events have focal depths between 35 and 40 km and could have occurred in the uppermost mantle (Suárez et al., 1981), as in the Tien Shan. There is no record of earthquakes with large magnitudes ($M > 7 \frac{1}{2}$) in the sub-Andes, and therefore individual faults are probably quite short. Fault plane

solutions indicate primarily thrust faulting with east-west trending P axes (Figure 6; Suárez et al., 1981). The dips of the planes are generally between 30° and 60° (Figure 2), with the shallower dipping plane usually dipping west (Figure 6). We presume that in most cases the west dipping plane is the fault plane so that the Brazilian shield underthrusts the Andes to the west. Most of the mapped faults on the east side of the Andes dip west (e.g., Audebaud et al., 1973), but there is no direct evidence that any of these faults were activated by the earthquakes that have been studied. In any case, a summation of the seismic moments shows that if the present rate of seismicity were typical of the last 80 m.y., the Andes could have been built solely by thrust faulting and crustal shortening (Suárez et al., 1981).

The rather steep dips of the west dipping planes from the fault plane solutions and the depths of these events suggests that the basement is involved in the deformation and that low angle decollement as in the Canadian Rockies in the late Cretaceous is not occurring, at least in this portion of the Andes. If the tectonics of western North America in the late Cretaceous and early Tertiary are similar to those of the Peruvian Andes now, then the deformation in the sub-Andes seems to be more similar to that of the Rocky Mountains in the early Tertiary (Laramide) than to that of the Sevier belt or Canadian Rockies in the late Cretaceous.

The active tectonics of the high Andes, as revealed by studies of Quaternary faulting and folding, show greater

variability than that in the sub-Andes. In central Peru, Quaternary gravels are folded in response to east-west or northeast-southwest shortening (Dollfus and Mégard, 1968; Mégard, 1978). Moreover, the fault plane solutions and surface deformation associated with the 1969 Parianhuanca earthquakes show reverse faulting, with a northeast-southwest trending P axis (Philip and Mégard, 1977; Stauder, 1975; Suárez et al., 1981). This sequence of events is unusual both because of the steep dip of the fault plane and because it occurred in a region of high altitudes. Most of the large events on the flanks of the Andes occurred beneath regions with elevations less than approximately 1000 m.

Whereas the high Andes of central Peru seems to be experiencing crustal shortening perpendicular to the range, north and south of this area there is evidence of normal faulting and crustal extension perpendicular to the range. This evidence includes observations of recent fault scarps that cut glacial moraines (Dalmayrac, 1974; Dalmayrac and Molnar, 1981; Mercier, 1981; Suárez et al., 1981) and of both surface faulting and the fault plane solution of the 1946 Ancash earthquake (Hodgson and Bremner, 1953; Richter, 1958; Suárez et al., 1981). We doubt that the normal faulting involves a large amount of extension, but it does imply a different stress distribution from that in the central Andes at high altitudes or that in the sub-Andes (Dalmayrac and Molnar, 1981).

Thus, the seismicity of the Andes is in some ways similar to that of the Tien Shan in that it indicates crustal shortening

and basement involvement. It differs by being bounded by a subduction zone and by the presence of normal faulting in some parts of the belt.

The Northwestern Himalaya, the Pamir, and the Hindu Kush

The discussion above is meant to present a sequence from a linear mountain belt with simple underthrusting of one continent beneath it (the Himalaya) to another linear belt, with deformation of the basement instead of simple underthrusting (the Zagros), to a two-sided linear belt experiencing crustal shortening and thickening with basement involvement (the Tien Shan) to another linear belt (the Andes) similar in some ways to the last but with greater variation along strike. In this respect the northwestern Himalaya, Hindu Kush and Pamir represent a further step from the simplicity of the Himalaya. Although linear ranges can be identified within the region, on the whole it shows little linearity. In parts it seems to be two-sided, with northward underthrusting beneath the Himalaya and southward underthrusting beneath the northern edge of the Pamir (Figure 7). Yet with strike-slip faulting and even some normal faulting, and with inclined zones of intermediate depth earthquakes, the region is clearly more complex than the others discussed above. Our purpose here is not to ignore or minimize this complexity but to isolate aspects of it and discuss them separately.

First, the discussions of the seismicity of the regions described above are based primarily on the study of teleseismically located events. Much of the key information comes

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from the study of a small number of events: the largest events since 1962 when the WWSSN began operating. Few such shallow earthquakes have occurred in the northwestern Himalaya, Hindu Kush or Pamir and most of what we know about the seismicity of this region is provided by studies of small earthquakes or microearthquakes using local networks. We suspect that similar studies of other areas will reveal complexities not apparent with only teleseismic data, but nevertheless we doubt that the difference in complexity for instance between the Himalaya in Nepal and in Pakistan will vanish with more data from studies of microearthquakes.

Studies of earthquake sequences or microearthquakes in the northern Pamir (Jackson et al., 1979; Kieth et al., 1981; Wesson et al., 1976), in northern Pakistan (Armbruster et al., 1978, Jacob et al., 1979; Seeber and Armbruster, 1979; Seeber and Jacob, 1977), and in northeastern Afghanistan (Prevot et al., 1980) all show a spectrum of fault types and orientations. In all three areas, the crust seems to experience complicated deformation on planes with different strikes and dips. Although there clearly are major strike-slip faults, which accommodate large displacements, they are not obvious from the seismicity and are either aseismic or have been quiescent for their recorded history. Except for a large event in 1505 near Kabul (Heuckroth and Karim, 1970; Quittmeyer and Jacob, 1979), there is little seismic activity on the portions of the major strike-slip faults in this area - the Chaman (Lawrence and Yeats, 1979; Yeats et al., 1979; Wellman, 1966),

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Darvaz-Karakul (Kuchai and Trifonev, 1977), Aksu-Murgab (Ruzhentsev, 1963), and other faults. There may also be major thrust (or normal) faults that are not defined by the seismicity. Instead, the seismicity scatters throughout much of the area, defining few linear trends. Yet within this complexity there appears to be a consistent north-south to north-northwest - south-southeast orientation of P axes of fault plane solutions. Despite some exceptions, this pattern suggests that the crust is undergoing north-south or north-northwest - south-southeast crustal shortening, presumably in response to the convergence of India and Eurasia. The variety of fault plane solutions and the widespread seismicity implies more homogeneous deformation than in the regions discussed above. We cannot exclude the possibility that a longer history of seismicity will reveal concentrated deformation on faults that bound relatively aseismic blocks. Nevertheless, it is clear that the tectonics of this region cannot be described easily by plate tectonics, unless one wants to use tens or hundreds of small plates.

The Hindu Kush and Pamir are particularly unusual in that they are underlain by a belt of intermediate depth events. We think that in fact there are two zones: one dipping north under the Hindu Kush and the other dipping south or southeast beneath the Pamir (Chatelain et al., 1980; Roecker et al., 1980). Seismicity extends to 300 km depth, and between about 150 km and 300 km it is confined to a rather narrow zone (Billington et al., 1977; Chatelain et al., 1980; Roecker

et al., 1980). The fault plane solutions show some variability, but for events deeper than about 150 km, the T axes lie in the plane of seismicity and plunge steeply. This similarity with seismicity at island arcs suggests that within the last 10 to 20 m.y., oceanic lithosphere was subducted beneath the Hindu-Kush and Pamir. Chatelain et al. (1980) inferred that two separate basins were subducted, one northwards beneath the Hindu-Kush and the other southward beneath the Pamir, and that one or both may have been interarc basins, like those in the western Pacific.

The seismicity between 70 and about 150 km depth shows some peculiarities atypical of subduction zones. First, seismicity shallower than about 70 km, within the thickened crust, is very low. This suggests that if deformation occurs in the crust, it is aseismic. The seismic zone between 70 and 150 km depth occurs over a broader zone and the fault plane solutions are much more variable than at greater depths. In general these fault plane solutions are consistent with north-south shortening of the region, but neither is the T axis consistently parallel to the dip direction of the seismic zone nor is the seismicity aligned with one of the nodal planes for each of the solutions. In any case, Roecker's (1981) inference that these earthquakes occur in subducted continental crust, possibly modified by high pressure phase changes, renders peculiarities in the seismicity a likely occurrence, even if we cannot at this time predict the nature of the peculiarities.

Thus, the seismicity of the northwest Himalaya, Pamir and Hindu-Kush differs in two ways from that of the regions discussed above. First, it is yet more complicated, with a mixture of thrust, strike-slip and even normal faulting and with diffuse deformation over a broad area. Second, the presence of inclined zones of intermediate depth seismicity in an intracontinental region suggests that there has been recent subduction of oceanic lithosphere. Although there may be some temptation to relate these differences to one another, we suspect that they are unrelated. The complicated deformation of the northwest Himalaya, Pamir and Hindu-Kush region make it resemble a miniature example of the broader zones between converging continents, such as in the vast region east of it, where intermediate depth seismicity is rare.

INTRACONTINENTAL CONVERGENT ZONES ON A LARGE SCALE

Most of the belts described are only portions of broad zones of deformation between converging continental plates. Whereas most, if not all, of the deformation occurring in these broad zones seems to be due to such convergence, the deformation is often more complicated and more varied than that described above for the individual ranges within these zones. Several factors contribute to this complexity. One is the inadequacy of the historic record of seismicity. To discuss consistent patterns in the tectonics requires a combination of seismic data with those from different sources

relating to Quaternary faulting. Together these data allow a qualitative description of the tectonics. Unlike in oceanic area, however, where seismicity occurs on narrow belts (plate boundaries) for which estimates of average rates of slip can be made, in continents the earthquakes and recent faulting are often dispersed over a broad area. Correspondingly deformation is not concentrated in a small number of intercontinental narrow zones, and it is difficult to determine rates of deformation.

Faulting in continental areas often includes a spectrum of fault types and orientations. In some, and perhaps in most cases, active faults develop along older zones of weakness. These inherited zones of weakness introduce a second important source of complexity to the active tectonics of continental regions generally absent in oceanic areas. Deformation in continents is diffuse, but zones of concentrated deformation are not completely absent. In particular, movement along large strike-slip faults seems to accommodate an important fraction of the convergence both between Arabia and Eurasia (McKenzie, 1972) and between India and Eurasia (Molnar and Tapponnier, 1975; Tapponnier and Molnar, 1976, 1977, 1979). Parts of many of these strike-slip faults seem to follow older faults, sutures, or other zones of intense deformation. Accordingly, whereas the deformation of continents is probably better approximated by the response of a continuum to external stresses than by the relative motion of rigid blocks (e.g., Tapponnier and Molnar, 1976), only some aspects can be approximated by a continuum with uniform properties.

Despite the complications due to inherited zones of weakness, large areas between converging continents often seem to be deforming in response to a relatively uniform stress field. Usually the P axes of fault plane solutions show a uniform orientation over a large area, and when they do not, usually the T axes are consistent. These axes are generally consistent with the direction either of crustal shortening or of crustal extension implied by the orientations of folds and faults and by the senses of motion on the faults. The consistency of these data imply a coherent regional stress field over large areas. |

In general the inferred regional stress fields change gradually over a broad region, but not always. In particular where there are large changes in mean elevation and in crustal thickness, the vertical stress can change from the maximum to the minimum compressive stress. Consequently, the concept of a regional stress field is limited in applicability or must be used cautiously.

To illustrate how these factors lead to complexity in the tectonics of continents, let us consider central and eastern Asia between India and Siberia.

Above we discussed the underthrusting of India and the creation of the Himalaya. Let us discuss Tibet below. North of Tibet, the Tien Shan experience crustal shortening, by thrust and strike-slip faulting, in an approximately north-south direction, parallel to the convergence between India and Eurasia. The seismicity in the last 20 years has been very low in the Nan Shan, northeast of Tibet (Figure 7), but studies

of Quaternary faulting indicate northeast-southwest shortening (e.g., Tapponnier and Molnar, 1977). East of Tibet, the predominance of thrust faulting seems to give way to strike-slip faulting, consistent with east-west crustal shortening. This rotation of the orientations of the P axes of fault plane solutions and directions of crustal shortening seems to be a characteristic feature of continental deformation. Such a rotation is predicted by calculated stress fields for continuous media indented by or stressed on particular boundaries (e.g., England and McKenzie, 1981; Tapponnier and Molnar, 1976), and it militates against interpretations in terms of a small number of rigid plates.

This gradual change in orientations of P, T, and B axes of fault plane solutions and in the orientation of the axes of principal strain (or stress) can also be seen further northeast. The predominance of thrust faulting in the Tien Shan and Nan Shan seems to pass into predominantly conjugate strike-slip faulting with northeast-southwest P axes and northwest-southeast T axes in Mongolia and north-central China (Figure 7). Further northeast and east, a large component of normal faulting dominates the strike-slip component in the Baikal Rift System, the Shansi Graben, and the region between them. The gradual change in the style of deformation and the orientation of strain implies that the stress field changes gradually across the region. The widespread deformation does not allow well-defined plate boundaries to be drawn except in restricted areas.

The seismicity of Mongolia illustrates the inadequacy of the historic record of seismicity. Fault plane solutions of earthquakes in the last 18 years (3-4 solutions) are wholly inadequate to justify the inference of important strike-slip faulting. Yet the surface deformation associated with several major earthquakes and Quaternary faulting seen both on the ground (Tikhonov, 1974) and on the Landsat imagery (Tapponnier and Molnar, 1979) demonstrate large-scale conjugate faulting. In particular, two major events ($M \sim 8.5$) in 1905 in northern Mongolia ruptured an east-west zone 370 km long causing several meters of left lateral slip (Florensov and Solonenko, 1965; S.D. Khilko, personal communication, 1974). In 1957, another major earthquake ($M \sim 8.3$) ruptured a 270 km long east-west zone in southern Mongolia, again causing several meters of left lateral motion (Florensov and Solonenko, 1965). Then in 1967, a smaller earthquake ($M \sim 7 \frac{3}{4}$) caused between 1 and 2 meters of right lateral slip along a 40 km fault that trends north-south (Natsag-Yam et al., 1971). Moreover, in 1931 a major event in China just west of Mongolia caused several meters of right lateral slip in a north-northwest - south-southeast trending fault (Deng Qidong, personal communication, 1981). Thus, although the seismicity in the last 20 years has been too low to define a pattern, the large earthquakes, which ruptured long faults, corroborate the inference of conjugate faulting seen on the satellite imagery (Tapponnier and Molnar, 1979).

The Baikal Rift Zone and the Shansi Graben also illustrate the difficulties with the historic seismic record. The Baikal system is clearly active, both with large events and microearthquakes (e.g., Golonetskii, 1975). Fault plane solutions of earthquakes with $M \sim 6$ (Misharina, 1967; Tapponnier and Molnar, 1979) and with $M \sim 4-5$ (Misharina and Solonenko, 1972), as well as surface faulting associated with recent earthquakes (Solonenko et al., 1966a, 1966b, 1968) attest to active normal faulting. Seismicity in the Shansi Graben, however, has been essentially non-existent since the turn of the century (Lee et al., 1978). There are no fault plane solutions for earthquakes within it, although there are for several events to the east. Yet the Chinese catalogue of earthquakes lists 2 events with $M \sim 8$ and another for $M \sim 7$ for this region, prior to 1800 (e.g., Lee et al., 1976). One must conclude that the graben is active, even if there has been negligible seismicity for nearly 200 years.

These portions of eastern Asia illustrate several different styles of deformation, but we think that all of them can be related to the penetration of India into Eurasia. This penetration causes thrust faulting and the creation of high mountains near the suture zone and strike-slip and normal faulting further from the suture. Crudely India seems to squash (and to have squashed) Eurasia directly in front of it and further away to wedge it apart. Nevertheless, if the seismicity and active deformation of Mongolia, Baikal and Shansi seem only indirectly related to the collision between India

and Eurasia, there is one other portion of eastern Asia in which this relationship is even less apparent - Tibet.

Although Tibet lies directly in front of India, at present it experiences normal faulting and east-west extension instead of north-south shortening and thrusting (Chen et al., 1981; Molnar and Tapponnier, 1975; 1978; Ni and York, 1978). Some 15 fault plane solutions, including one for an event at 90 km depth (Chen et al., 1981), show this extension (Figures 1 and 7), which is corroborated both by the analysis of satellite imagery (Molnar and Tapponnier, 1975; 1978; Ni and York, 1978) and by observations made on the ground (e.g., Bally et al., 1980). Note that this style of deformation calls for a major change in the "regional stress field" from that in the Himalaya or Nan Shan and Tien Shan. We relate this extension and change in "regional stress field" to the gravitational buoyancy forces acting on the high elevation and thick crustal root of Tibet (Molnar and Tapponnier, 1978). To balance the gravitational body force, either horizontal compressive stresses applied in the surrounding areas of low elevations or the strength of the elevated (and thickened) crust must hold the area together, preventing the collapse and spreading apart of the mountains. Apparently India's convergence with Eurasia provides the necessary north-south compressive stress, but neither the east-west compressive stress on the flanks of Tibet nor the strength of Tibet are adequate to prevent it from extending in an east-west direction.

We presume that this extension began recently and that the amount of extension is quite small. At the same time the

variation in the direction of underthrusting from northeast in the northwestern Himalaya to north-northwest in the eastern Himalaya implies a rate of extension across Tibet of about 1 cm/yr (Armbruster et al., 1981). In 5 to 10 m.y., extension would amount to 50 to 100 km, which would indicate a small amount of strain ($\leq 5\%$) within the plateau compared with that which probably occurred during crustal thickening ($\sim 100\%$).

The active tectonics of eastern Asia illustrate a variety of fault types and fault orientations. The pattern of deformation is not simple at any scale. Yet, we think that all of this deformation can be related to India's collision with and penetration into Eurasia. In some regions the relationship is simple: India pushes on Eurasia and causes crustal shortening and thrust faulting. In other areas, relatively strong blocks are pushed as rigid bodies, causing deformation on their edges. The eastward motion of such blocks with respect to Eurasia probably contributes in part to the opening of the Baikal rift system and the Shansi Graben zone. The relationship to the collision is even less direct in Tibet. There the present strain field and style of tectonics owe their existence more to the previous tectonic history, during which the plateau was formed probably by north-south crustal shortening, than to the continued penetration of India into Eurasia. This less direct relationship suggests that during the evolution of a belt, the seismicity and active tectonics of a mountain belt can pass through phases as crustal shortening causes crustal thickening and increases the vertical normal

stress that in turn resists further thrust faulting (Tapponnier and Molnar, 1976).

DEPTHS OF FOCUS

We separate the discussion of depths of focus in order to call attention to what appears to be a simple pattern: in continental regions, seismicity usually is confined to the shallower part of the crust but occasionally occurs also in the uppermost mantle, just below the Moho. The lowermost crust, however, in general seems to be aseismic (see Chen and Molnar, 1981a, 1981b, for more details).

Most intermediate and deep focus earthquakes occur at subduction zones and are assumed to occur in downgoing slabs of lithosphere. Those that occur in continental settings are often attributed to recent subduction of oceanic lithosphere. It is now becoming clear that there are intermediate depth events* that cannot be associated with subduction of oceanic lithosphere and that there are numerous other shallower, sub-crustal events that also cannot. Chen and Molnar (1981a, 1981b) compiled a list of depths of well-located events not associated with subduction zones and found the majority to be in the upper 15 km. In stable shields and platforms depths as great as 25 km were noted. Greater depths also were encountered where thrust faulting and crustal shortening actively occur. In those regions, the colder temperatures of the shallow crust

* Strictly speaking, intermediate depths are usually defined to be between 70 and 300 km.

are transported down by underthrusting. In a few areas, earthquakes have been precisely located in the mantle at depths from 45 km to 100 km, and apparently even 150 km in the High Atlas Mountains (Hatzfeld and Frogneux, 1981). Yet, in general, the lower crust, however thick the crust may be, seems to be devoid of seismicity.

These observations can be taken to indicate the following: Earthquakes in the crust occur in material with temperature less than about 300°C ($\pm 100^{\circ}\text{C}$). Crustal material with higher temperature probably flows and does not fracture suddenly. Earthquakes in the mantle, however, probably occur where the material is colder than about 800°C ($\pm 200^{\circ}\text{C}$) (Chen and Molnar, 1981a, 1981b; Chen et al., 1981; Molnar et al., 1979). Olivine is probably much stronger than most crustal minerals at the same temperature, and there may be a low strength zone in the lower crust (although not necessarily at the Moho) (Figure 8). Crystalline nappes are likely to detach within this low strength zone.

SUMMARY

Unlike island arc structures, the seismicity of mountain belts exhibits considerable variability. There is no typical mountain belt exemplifying characteristics common to all other belts. Instead, different belts exhibit different styles of deformation that probably result from different earlier tectonic histories and from different stages of evolution in their present development.

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The seismicity of the Himalaya implies that India is sliding beneath the Himalaya along a shallow dipping fault near, if not along, the top surface of the Indian Shield. The seismicity gives no evidence of active detachment of a crystalline nappe from the top of the Indian Shield, but sediments deposited in the Ganga Basin could be experiencing decollement (Seeber et al., 1981; Seeber and Armbruster, 1981).

The seismicity of the Zagros is different in that at present there is no clear seismic evidence for decollement. Instead the earthquakes seem to occur in the basement beneath the thick evaporite and carbonate layers. The basement seems to be deforming not by low-angle faulting but along steeper faults (30-60°). Jackson (1980b) suggests that ancient normal faults that formed in an earlier rifting phase are reactivated as reverse faults.

The Tien Shan is a broader belt with active seismicity showing thrust faulting throughout. The belt is two-sided. On the north the Siberian platform underthrusts southwards and on the south the Tarim basin underthrusts northwards. Thrust faults seem to dip steeply (30-60°), so that simple decollement, or detachment of thin crystalline sheets does not seem to be occurring unless the faults change dip at greater depths. In addition, there are prominent strike-slip faults in the Tien Shan for which there is no seismic evidence for recent activity, but which show clear evidence of movement in the Quaternary. The sense of strike-slip movement is consistent with a north-northwest - south-southeast maximum compressive stress inferred from the fault plane solutions.

Westward underthrusting of the west Brazilian craton beneath the Andes also occurs on faults dipping steeply (30-45°) into the basement. Depths of focus require that the larger events occur within the basement, not in the sedimentary cover. Thus the active tectonics more closely resembles that of the western United States during the early Tertiary (Laramide) than that of the Sevier belt or the Canadian Rockies in the late Cretaceous. The active tectonics of the high Andes shows either normal or reverse faulting, depending upon the locality along the zone.

The seismicity of the northwest Himalaya, Pamir and Hindu-Kush shows greater variety than that of the other belts. Although strike-slip faults, clearly active during the Quaternary, bound the region on its east and west, and although clear thrust faulting occurs on the northern and southern edges of the region, the seismicity suggests diffuse deformation with a spectrum of fault plane solutions.

The northwest Himalaya, Pamir and Hindu-Kush region is in many ways a microcosm of the broader region further east or west between Eurasia and India or Arabia. In these regions, deformation also includes a spectrum of fault types and fault orientations. Strike-slip faulting, in particular, seems to play a crucial role in displacing material and allowing convergence of continental masses to continue without building mountains with limitless elevations. In general the orientation and sense of motion on the strike-slip faults is consistent with a regional stress field resulting from convergence between the major continental plates. More

problematic is the evidence for normal faulting in portions of these broad convergent zones. We think that this extension nevertheless is attributable to the collision process. Some normal faults may result from buoyancy forces acting on regions of high elevation and compensating crustal roots (Tibet) and others (Baikal and Shansi Grabens) may result from wedging apart of Asia by movement of relatively rigid blocks past one another.

Several factors contribute to the greater complexity in continental convergent zones than at subduction zones. One factor is the inadequacy of the seismic record. Whereas the rate of convergence at subduction zones is usually several cm/yr, rates of mm/yr are likely to characterize individual fault zones in a continental region. Thus, seismicity alone is likely to be lower on such continental fault zones than on plate boundaries in oceanic regions. In addition, the role of fault creep is clearly important in some continental regions, and in others (like Mongolia) a substantial amount of deformation occurs during very large earthquakes. This variation both in the frequency and the maximum size of earthquakes and in the importance of fault creep from region to region is an obvious but poorly understood phenomenon in continental seismicity.

A second factor contributing to the complexity of continental seismicity is the variation in strength of the earth's crust in continents compared with the more homogeneous oceanic lithosphere. Variations in strength result both from

differences in temperature profiles (e.g., Molnar and Tapponnier, 1981) and from inherited zones of weakness. For instance, the major strike-slip faults of the Tien Shan seem to have been active in the late Paleozoic, and the existence of large relatively rigid blocks, like the Tarim Basin, cause a concentration of deformation on their margins. An eastward motion of cold, strong blocks relative to Eurasia may contribute to the opening of the Baikal and Shansi grabens. Thus the variation in strength both causes stress concentrations and alters the orientation of stress in continental regions.

Another phenomenon that alters the stress field is the variation in crustal thickness (e.g., Artyushkov, 1973; Dalmayrac and Molnar, 1981; Frank, 1972; Molnar and Tapponnier, 1978). In the equations of equilibrium the gravitational body force, which varies from regions of high and low elevation and of thicker and thinner crust, must be balanced by variations in the stress field. In particular, regions of thick crust are likely to experience horizontal extension. Surrounding regions of thinner crust and lower elevations are more prone to horizontal compression. Thus, differences in the style of deformation and in the seismicity are likely between regions of high and low elevation, as can be seen in Tibet and the neighboring Himalaya and Nan Shan or in portions of the Andes.

If the buoyancy of continental crust does perturb the stress field, one can imagine that during the history of a continental convergent zone, a mass of rock might first experience rapid thrust faulting while crustal thickening occurs, and then reverse faulting at a lower rate while the

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area being deformed grows in size. Later strike-slip or normal faulting might develop when the convergence or the movement of surrounding material change in some way so as to relax the horizontal compressive stress in one or all horizontal directions (Tapponnier and Molnar, 1976). The observed differences in the seismicity of different belts may exist partly because each is at a different stage in its development.

One other factor probably contributes to the greater complexity of continental seismicity than that of subduction zones. Laboratory studies show that olivine is much stronger than typical crustal rocks and minerals at the same temperature. Thus oceanic lithosphere, which consists mostly of mantle material, probably can withstand much larger stresses than continental lithosphere can. Continental lithosphere consists of thick crustal layer, and consequently does not contain the thick zone of cold mantle ($T < 500^\circ$) that characterizes oceanic lithosphere. A study of depths of earthquakes (Chen and Molnar, 1981a, 1981b) supports the inference that the continental lithosphere may contain a weak zone in lower crust. The lower crust appears to be aseismic, while most seismicity occurs at shallow depths in the crust. In some regions the mantle lithosphere is also seismically active. We infer that the lower crust deforms by aseismic (ductile) flow and may be weaker than the overlying brittle crust and underlying uppermost mantle. Crystalline nappes may detach along this zone in the lower crust.

Although the seismicity and tectonics of continental regions are more complicated than those of subduction zones, we think that this complexity owes its existence to relatively simple physical phenomena. A key problem for the future is to evaluate quantitatively the relative importance of these phenomena. Clearly a continued and even expanded interaction of geologists from different disciplines will be necessary to bring about a quantitative understanding of continental tectonics that is comparable to that of plate tectonics in oceanic regions. It is also clear that we seismologists still have many unresolved problems that must be addressed before we will have done our part.

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FIGURE CAPTIONS

Figure 1: Neotectonic map and fault plane solutions of earthquakes in the Himalaya. Lower hemisphere projections of fault plane solutions (from Armbruster et al., 1981; Molnar et al., 1977; Molnar and Tapponnier, 1978) of earthquakes between 1963 and 1976 which are large enough to study with the WWSSN are plotted with blackened quadrants containing compressional first arrivals. When two events occurred at nearly the same location, only one is shown. Black dots are epicenters of major historical earthquakes. Faults in the Himalaya are from Gansser (1977), and those in the Tibetan plateau are from Molnar and Tapponnier (1978).

Figure 2: Histograms showing dips of nodal planes for two subduction zones and four intracontinental mountain belts. Data are from Isacks et al. (1969) and Johnson and Molnar (1972) for Tonga-Kermadec, from Molnar and Sykes (1969) for Mexico and Central America, from Molnar et al. (1977) and Molnar and Tapponnier (1978) for the Himalaya, from Jackson and Fitch (1981) and McKenzie (1972) for the Zagros, from Tapponnier and Molnar (1979) for the Tien Shan, and from Suárez et al. (1981) for the Peruvian Andes.

Figure 3: Schematic cross-section across the Himalaya, showing an interpretation of the depths of earthquakes, fault plane solutions and active tectonics. Geology

Figure 3 (contd.):

simplified from Gansser's (1964) cross-section of the Kumaon Himalaya. Black dots indicate earthquake hypocenters with arrows showing the direction of relative motion along the fault planes.

Figure 4: Neotectonic map and fault plane solutions of earthquakes in the Zagros. Symbols as in Figure 1. Seismic data from Jackson and Fitch (1981) and McKenzie (1972). Geologic information from Ricou et al. (1977) and Geological Map of Iran (Geological Staff of the Iran Oil Co., 1957).

Figure 5: Neotectonic map and fault plane solutions of earthquakes in the Tien Shan. Symbols as in Figure 1. Seismic and geologic data from Tapponnier and Molnar (1979).

Figure 6: Fault plane solutions of earthquakes in the Peruvian Andes (from Suarez et al., 1981). Symbols as in Figure 1.

Figure 7: Recent faults, and fault plane solutions of Central and Eastern Asia. Epicenters of major historical earthquakes (large dots) and earthquakes with known fault plane solutions (small dots) are shown, together with orientations of the P - (—) and T - axis (\longleftrightarrow) for thrust and normal faulting events, slip vector (\rightarrow) of low angle thrusts, and relative motion along the fault plane (\rightleftarrows)

Figure 7 (contd.):

for strike slip events (dashed when one of the nodal planes is arbitrarily chosen as the fault plane).

Figure 8: A schematic diagram of the variation of the mechanical strength of the crust and uppermost mantle with respect to depth for cold (solid curve), intermediate (dashed curve) and warm (dotted curve) geotherms. The upper part of the curve is based on linear relationship between the shear stress and normal stress to represent stick-slip or brittle failure behavior. The strength in the lower crust and upper mantle is controlled by flow laws of crustal and mantle materials, respectively. The dashed curve is smoothed to indicate possible gradual changes of strength in the brittle-ductile transition zone and the crust-mantle boundary (from Chen and Molnar, 1981b).

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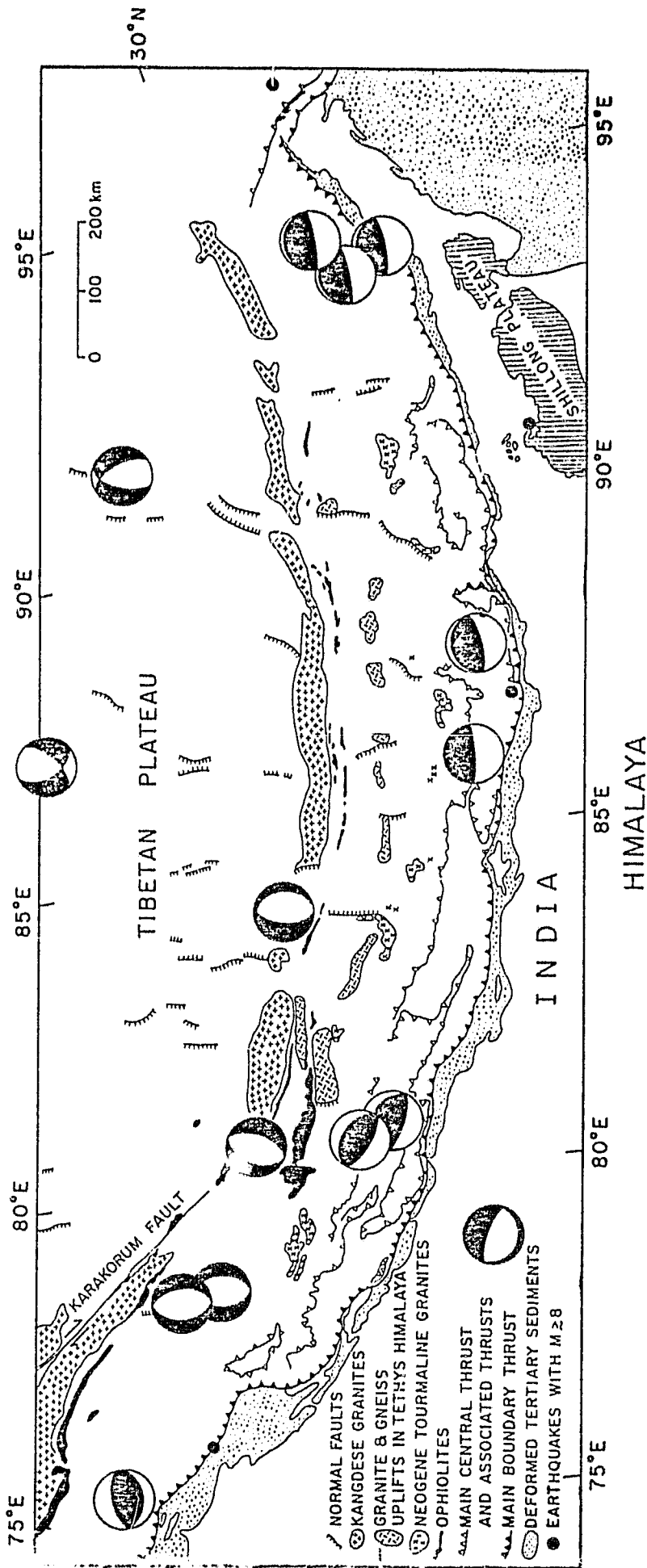
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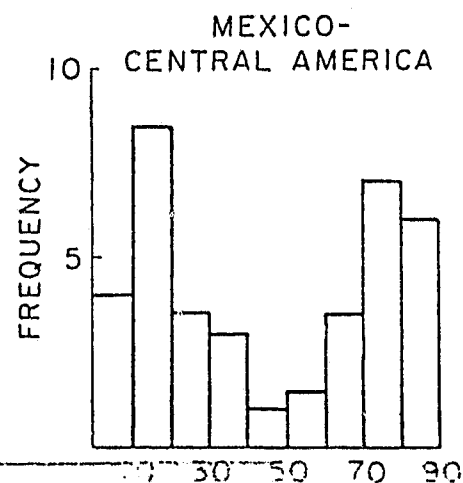
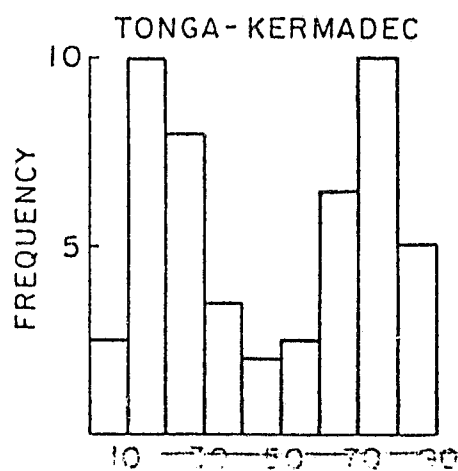
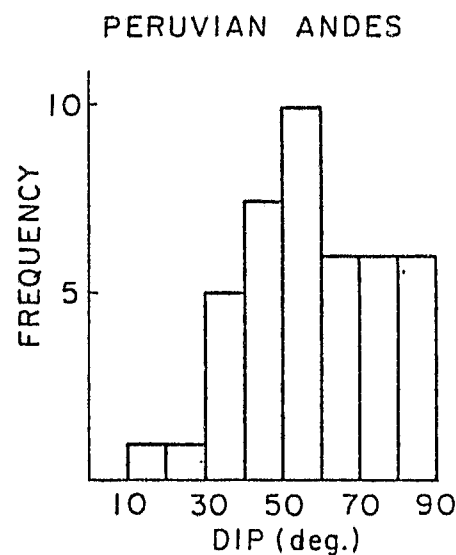
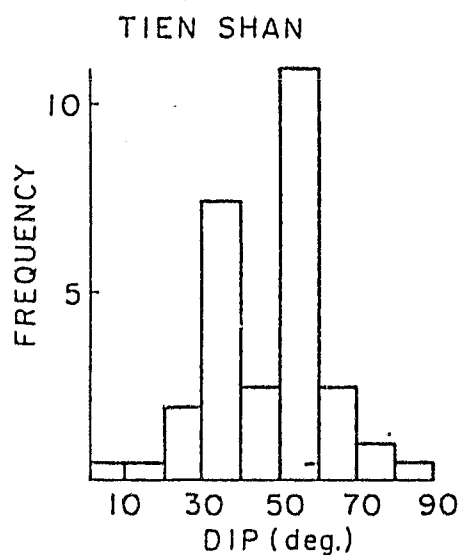
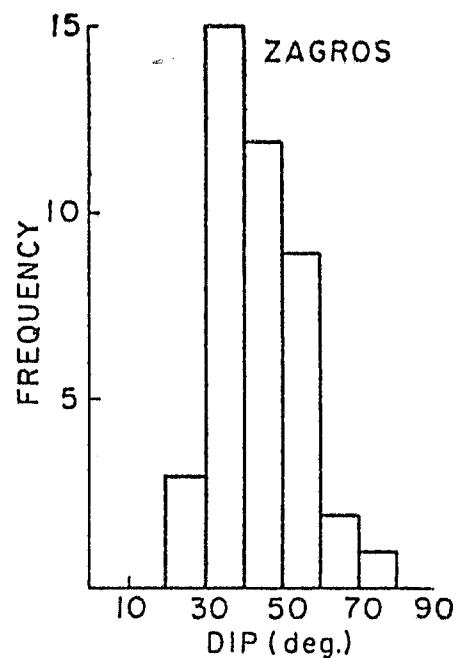
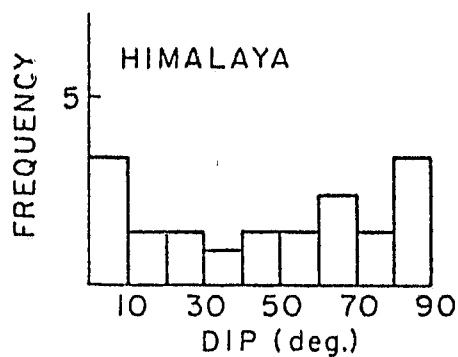
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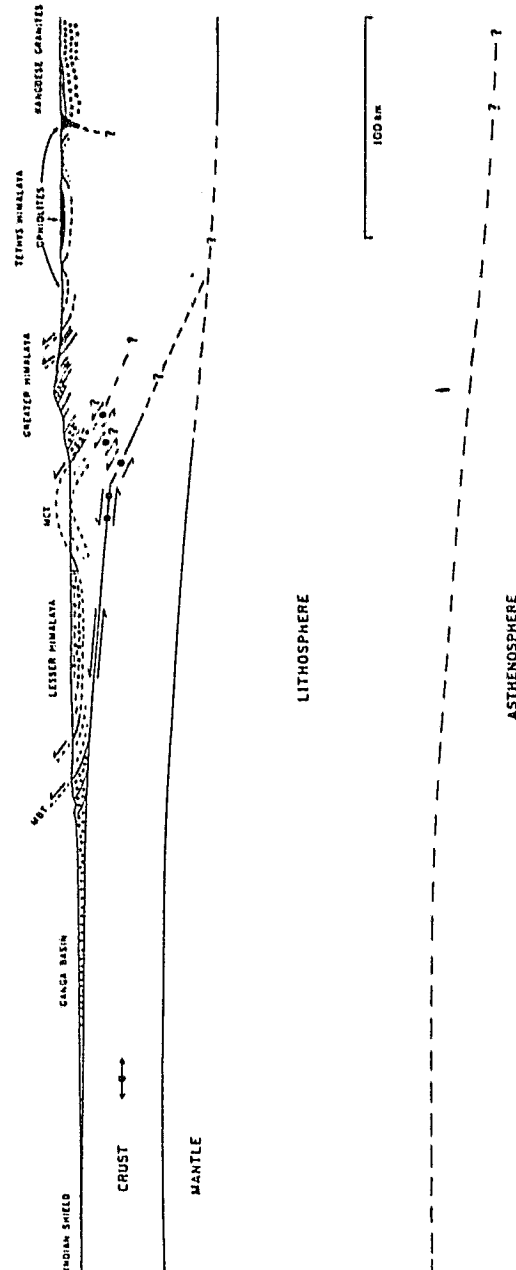
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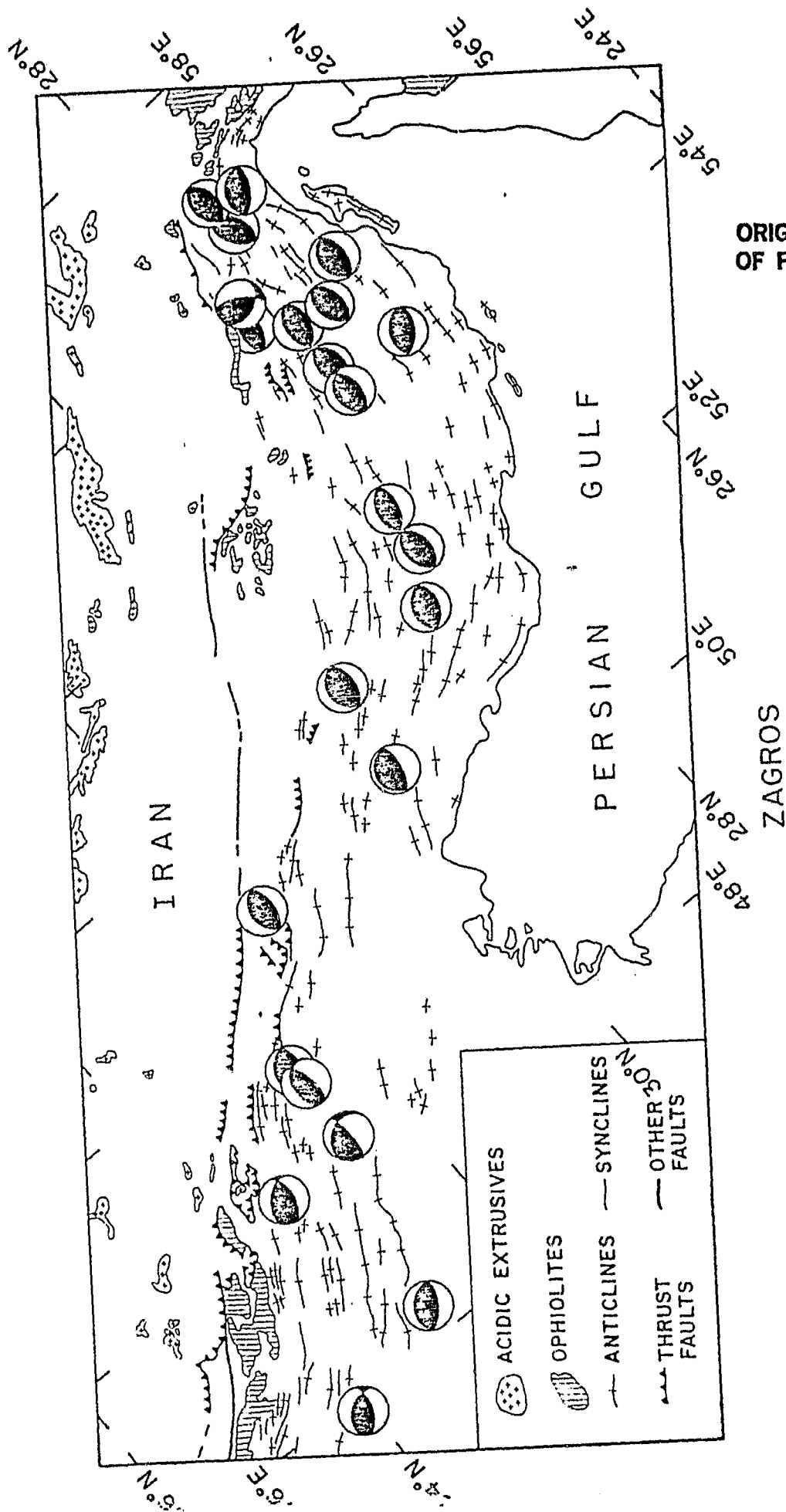


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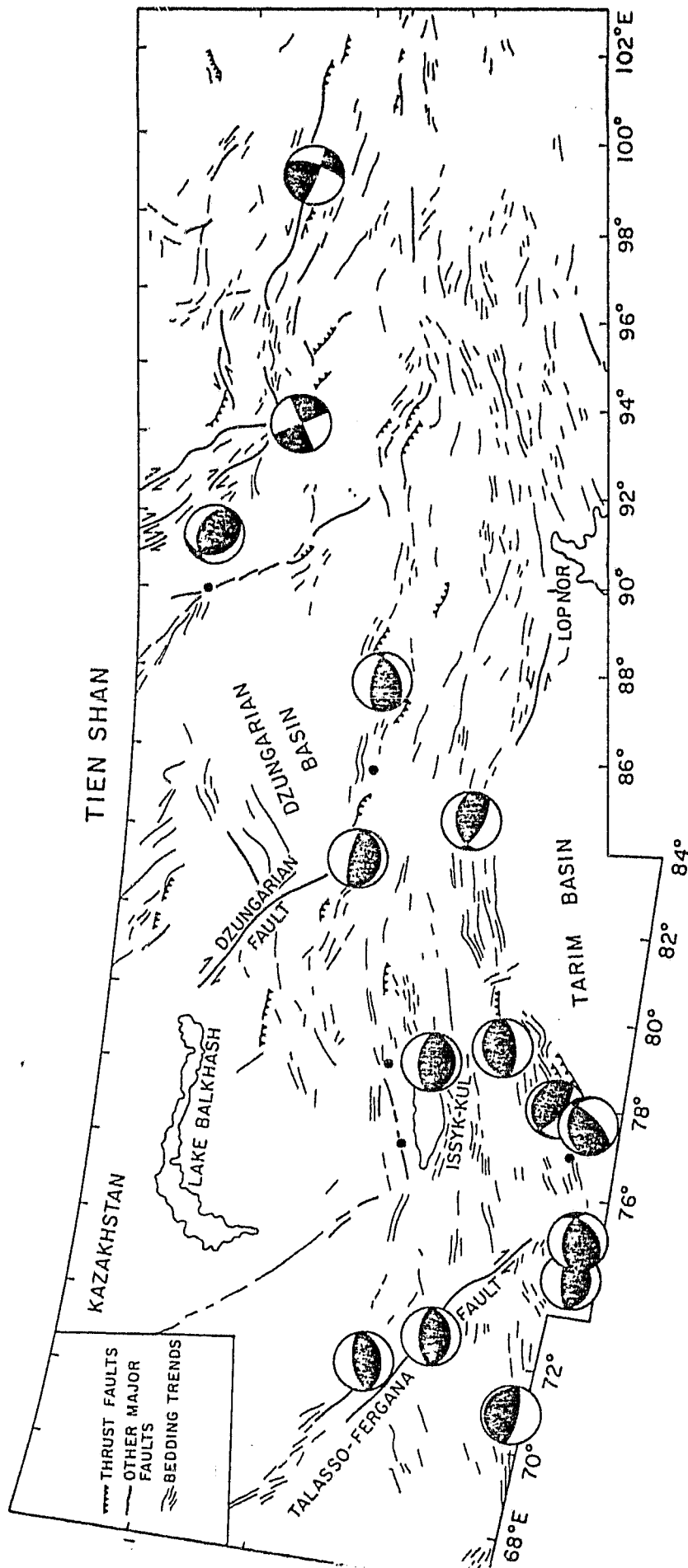


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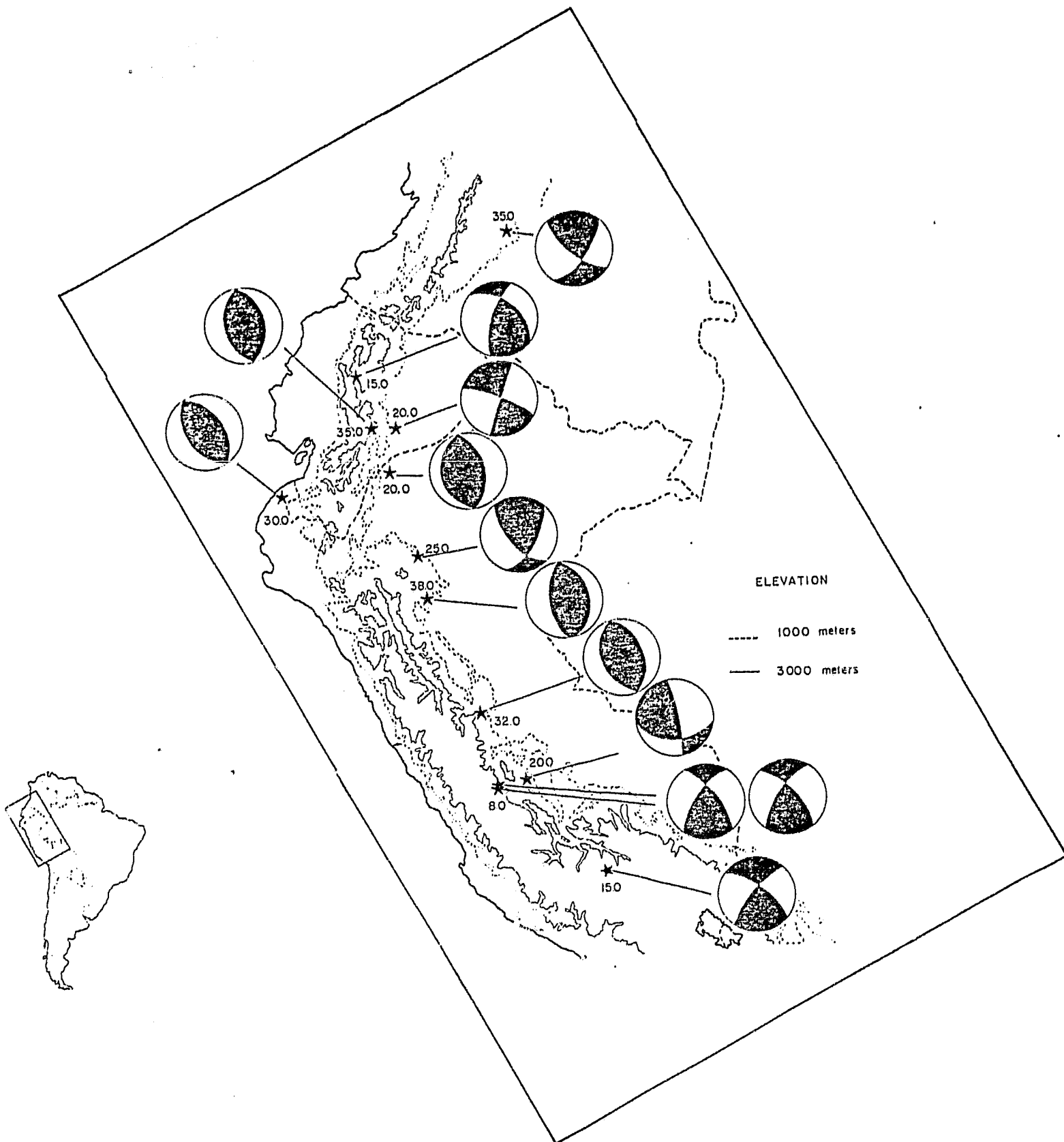


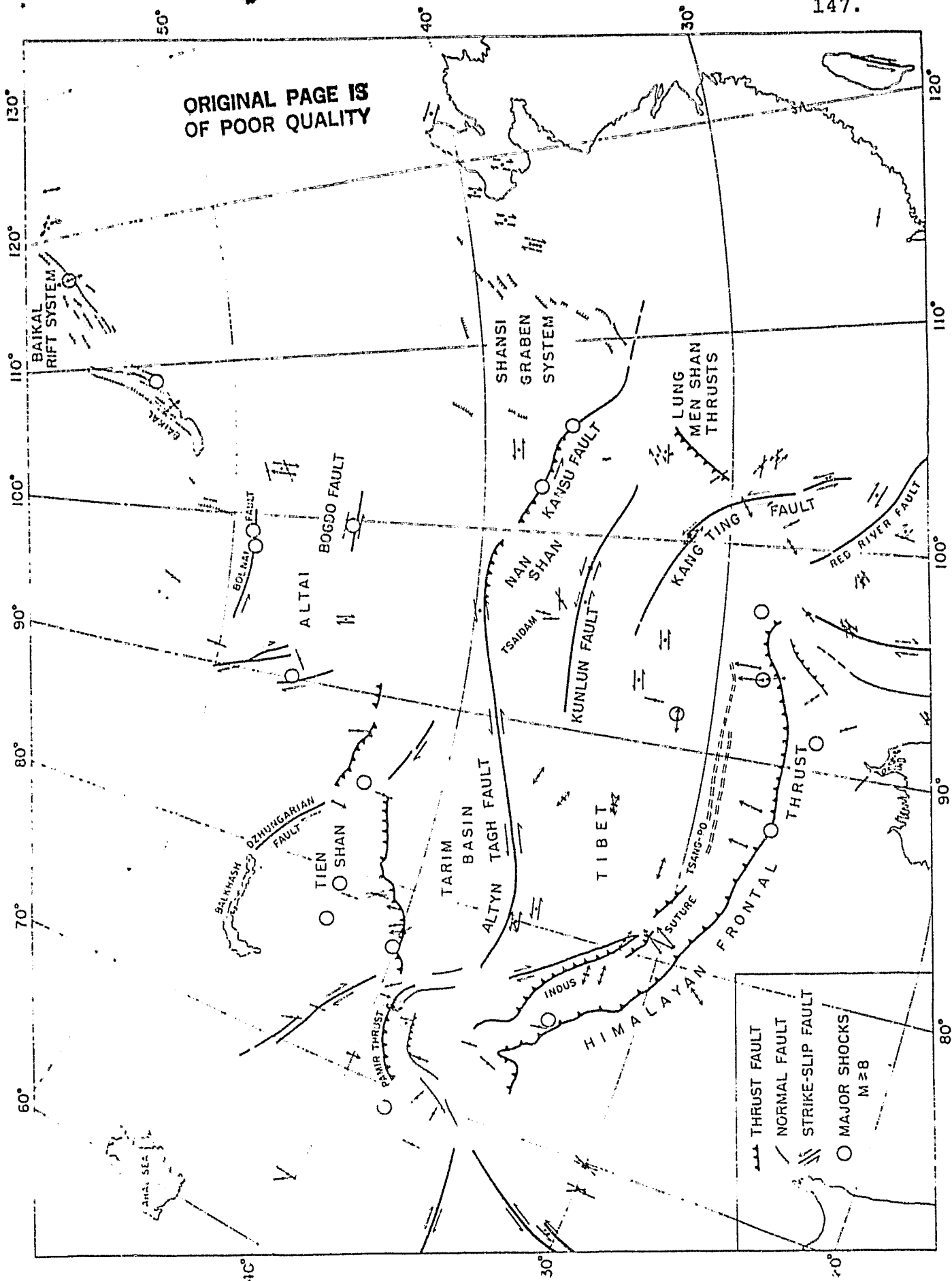
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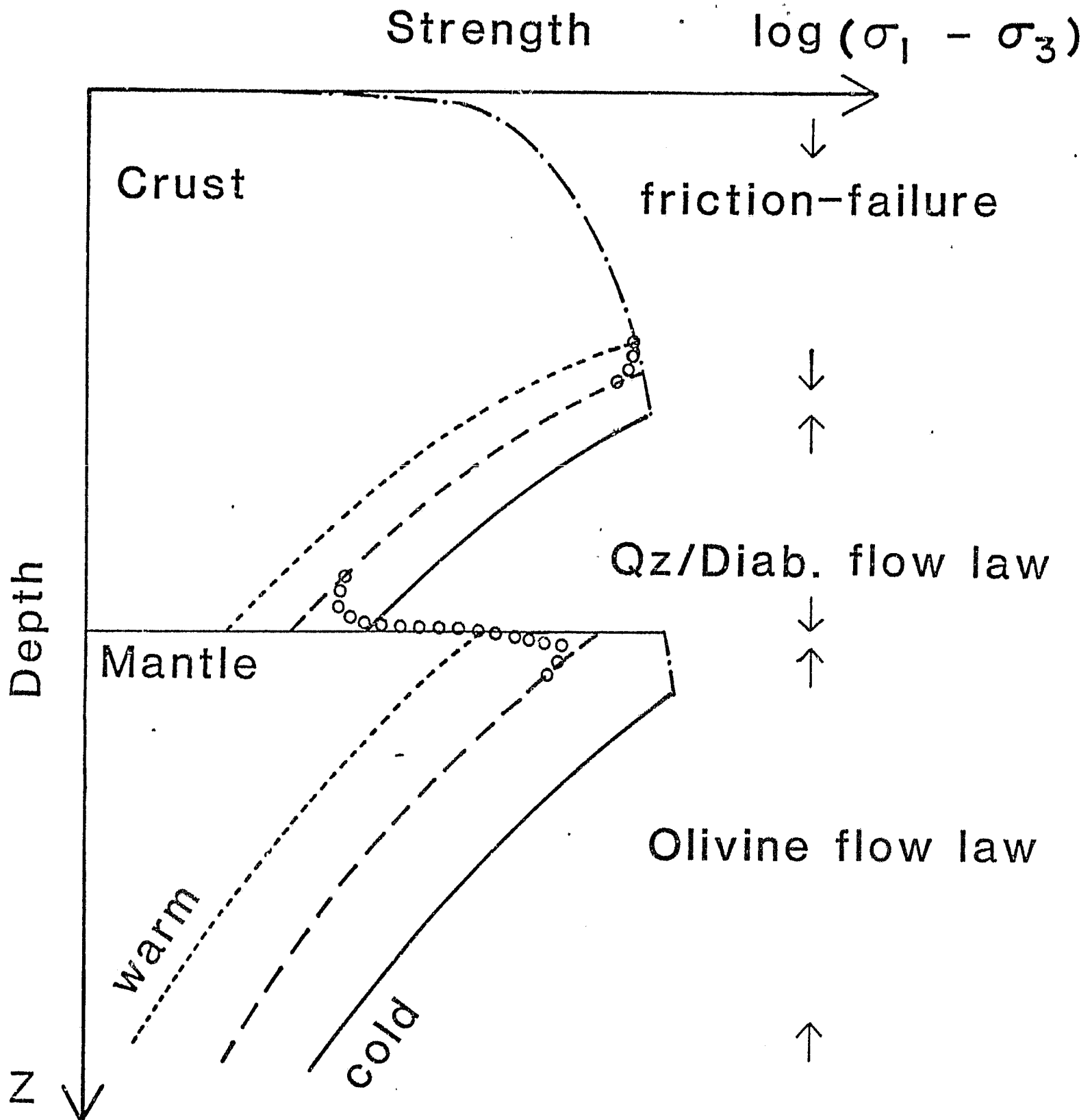
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APPENDIX A

METHODS, UNCERTAINTIES, AND PHILOSOPHY IN INTERPRETING SEISMIC RESULTS

Locations of earthquakes

Locating earthquakes reliably is a deceptively difficult process. Obtaining a location is relatively easy; usually one seeks a location and origin time that minimizes the differences in observed and calculated arrival times of P and sometimes S phases at various stations. Numerous computer programs for different station-source geometries exist and are routinely implemented. The major difficulty is estimating an uncertainty in the location, particularly in the depth of focus. The effects of different seismic wave velocity structures, station distributions, and uncertainties in arrival times on computed locations are often difficult to evaluate without simply making numerous tests with both real and synthetic data, with different velocity structures, and with various randomly added errors to the data. Some simple rules of thumb do exist, but we caution readers to pay careful attention to the discussions of uncertainties in published studies.

Both the U.S. Geological Survey (and formerly the U.S. Coast and Geodetic Survey and the National Oceanic and Atmospheric Administration) and the International Seismological Center routinely locate earthquakes using P wave arrival times reported by stations distributed throughout the world. During the last twenty years the number of stations used, the number

of earthquakes located, and the quality of the locations have all increased. These locations provide basic data sets that have been very useful for seismic studies of all kinds. Nevertheless, blind acceptance of these locations can lead to erroneous inferences. One must filter the reported locations by removing those with poor station distributions and with large residuals between observed and calculated arrival times. The uncertainties of the most reliable epicenters are probably at least 10 km, and systematic errors of as much as 20 km are known for well recorded and precisely located events (e.g., Cleary, 1967; Utsu, 1967). By careful examination of the distribution of stations used to locate the events and having small residuals, one can produce a list of events with uncertainties in epicenters generally less than 20 km. Such lists generally include only 1/3 to 1/4 of the locations reported by the USGS or ISC (e.g., Barazangi and Isacks, 1976, 1979; Billington et al., 1977; Isacks and Molnar, 1971; Molnar et al., 1973). The uncertainties in relative locations can be reduced by determining the locations relative to a nearby master event, but it is our opinion that relocations of individual events that occurred in the last 10 years using the same data that the USGS or ISC used are not likely to be significantly better than the locations reported by those agencies.

Depths of foci are considerably more uncertain than epicentral coordinates. Depths can be well constrained if phases reflected from the earth surface above the earthquake

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(pP and sP) are well-recorded. Such phases are often clear for events deeper than about 70 km but rarely so for shallower events. If there are stations closer to the epicenters than the focal depth, depths can be reliably obtained, but this is rarely the case for shallow events. Consequently all reported depths less than about 70 km for events located teleseismically should be viewed with suspicion, unless other information is used to support the reported depth.

Recently there has been considerable progress using synthetic seismograms to constrain depths of shallow events. Using the known fault plane solution one calculated theoretical seismograms for P or S phases (including sP and sP or pS and sS) for different depths and compares them with observed seismograms. The method works best for simple events, whose dimensions are sufficiently small and for which the slip on the fault is sufficiently abrupt that the source can be adequately modeled as point source. With synthetic seismograms, the uncertainties in depths can usually be reduced to less than 5 km.

For earthquakes recorded by local networks, locations can be more accurate than when only teleseismic data are used. For earthquakes occurring within the network, both epicenters and depths can be precise within a few km. Depths depend upon the distances to the nearest stations. Depths rapidly become unreliable the more the epicentral distance to the nearest station exceeds the focal depth. For events outside the network, uncertainties in both depths and epicenters can be very large,

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and generally we do not trust such locations unless special studies have been made to examine the uncertainties.

Once locations and their uncertainties have been determined, geologic inference can include a spectrum of seemingly conflicting possibilities. Whereas the occurrence of earthquakes attests to current tectonic activity, the absence of earthquakes can mean tectonic stability, merely temporary quiescence, or aseismic, but possibly rapid, deformation. Clearly geologic inference in the absence of other geologic information can be risky, if not simply subjective. An often asked question is whether a given distribution of earthquakes that occurred during a finite time interval is representative of the long term seismicity. It is our belief that in most regions thousands of years of data will be necessary to establish the long term average characteristics of the seismicity. If this is so, then one is safer posing questions for which a detailed knowledge of the long term seismicity is not important. Thus, while we can use seismicity to study aspects of the active tectonics of selected regions, we must recognize that there are both aseismic processes and discontinuous processes about which our short record of earthquakes can tell us little or nothing.

Fault plane solutions

Most fault plane solutions are determined using only the initial motions of P waves, but solutions can be improved using initial motions or polarizations of S waves, the complex

spectra or amplitudes of surface waves, or the wave forms of body waves. It is our opinion that the orientations of the nodal planes determined in fault plane solutions with P wave initial motions are rarely more certain than $\pm 10^\circ$. The other techniques mentioned above can improve the precision of the solutions, but because lateral variations in earth structure can alter the ray paths substantially and distort the solutions (e.g., Engdahl et al., 1977; Solomon and Julian, 1974), we still think that 10° is a safe lower bound for the uncertainty.

With teleseismic data, the most precisely determined solutions are those for strike-slip faulting on vertical planes. Both nodal planes are vertical and the data usually constrain them tightly. The poorest determined solutions are for dip slip faulting on planes dipping at 30° to 60° . Most of the data come from stations far enough away from the source and for which the take-off angles of the rays are steep so that first motions of only one sign cluster in the center of the focal sphere. There is no doubt about the type of faulting - normal or thrust - but the orientations of the nodal planes are not constrained. Moreover, because of both lateral variations in and an inadequate knowledge of the earth's upper mantle structure, the direction of P waves leaving the source to stations closer than about 30° is very uncertain. Unfortunately fault plane solutions of earthquakes in continents are often of this type and therefore are poorly constrained (Figure A1).

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When local networks are used, the uncertainties in the ray paths are even greater and can depend strongly on the assumed velocity structure. Whereas the location of an earthquake might be insensitive to whether rays leave the source upwards or are refracted by deeper layers before returning to the surface, the fault plane solutions depend critically on an accurate knowledge of the directions the rays leave the source. In some cases, particularly for strike-slip faulting, the solutions are not very sensitive to the assumed velocity structure, but in others, it can be impossible to determine a solution because of this dependence. Accordingly sometimes it is very difficult to estimate the uncertainty of the parameters describing a fault plane solution, because one cannot evaluate the uncertainties in the ray paths leaving the source. The best measure of the quality of the solution is probably given by the number of reliable readings used. Solutions based on only 8 or 10 first motions are less reliable than those based on 15 or 20. Often to increase the number of data, composite fault plane solutions are determined, using many different earthquakes. It is our opinion that composite fault plane solutions are useful only when the data from events in a small volume are very consistent with only one pair of nodal planes. We do not put much faith in composite solutions for which data from neighboring regions of diffuse seismicity give different solutions and with 10% or more of the readings inconsistent with the inferred solutions. One good solution for one event is more valuable than several composite solutions.

Moreover, solutions for bigger events are probably more representative of the large-scale deformation than those of small events.

For published discussions of fault plane solutions, the rules of thumb are (1) if a stereographic projection of the focal sphere is not shown or not readily available in the referenced literature, it is often safer to ignore the reported solution than to believe it; and (2) if the data are presented and plotted reliably, it is an easy matter for anyone familiar with stereographic projections to examine how consistent the P wave initial motions are and how well they define two orthogonal planes. When additional data (S waves, surface waves, etc.) are used, it can be more difficult for the non-seismologist to obtain an appreciation of the uncertainty in the solution, but in some cases the authors do address this question directly.

Once one has a fault plane solution there is some uncertainty in how to interpret it. The solution is defined by two nodal planes. One is the fault plane, and the normal to the other, the slip vector, gives the direction of relative motion on the fault plane. From the fault plane solution alone, one cannot choose which of the nodal planes is the fault plane. Additional information is needed to resolve this ambiguity - mapped faults, planes or belts of seismicity, etc. Sometimes the ambiguity cannot be resolved. Clearly, if we can identify the strike of the fault plane from other geologic information,

the fault plane solution can be used to give the dip of the plane and the direction of slip on it. But even when this is not the case, solutions often provide useful information about the active tectonics in the area.

Fault plane solutions can be uniquely described by the strikes and dips of the two nodal planes or by the plunges and azimuths of the normals to these planes. They also can be described by the orientations of the P, T, and B axes. The P and T axes lie at 45° to the two nodal planes in the quadrants with dilatational and compressional initial motions, respectively. The B axis is parallel to the line of intersection of the two planes. The P, B, and T axes crudely, but often inaccurately, approximate the directions of the maximum, intermediate and least compressive stresses. The solution can be described by any two of the planes or axes, and in fact requires only three angles for its complete description, the strike and dip of one nodal plane and the direction of slip in that plane. Thus, the axes and planes are not unique quantities. Which among them are physically important and which are merely consequences of the others are subjective questions whose answers are not universally agreed upon. We give some opinions below.

Along well-defined faults or fault zones such as plate boundaries, the important parameters seem to be the orientations of the planes and slip vectors. The P, T, and B axes do not seem to be quantities worthy of discussion. For instance where spreading centers and transform faults intersect, the

horizontal components of the slip vector do not change, but the T axis rotates 45° and P and B axes change by 90° . If the regional stress field changed by this much in such a short distance, then the concept of a regional stress field would not be very useful. Similarly the consistent orientation of the P axes for earthquakes along major strike-slip faults or at underthrust zones of island arcs is merely a consequence of the directions of motion along the faults. The stress field is likely to change along the fault, but because it is a zone of weakness, the fault controls the response of the region to the local stress field.

In regions with numerous faults, however, slip will occur on those with orientations most suitably oriented with respect to the stress field. In such a case, one can expect slip on nearby planes with different orientations. The slip vectors and nodal planes might scatter widely, but either the P or the T axis might be consistent among these solutions. For instance, in the Western Basin and Range province of North America, both strike-slip and normal faulting are common, but the T axes are usually oriented northwest-southeast. This suggests that the region is being stretched in a northwest-southeast direction.

The simple rule of thumb with fault plane solutions for many earthquakes in a region is that the consistent parameter is the important one. If the nodal planes are parallel, then probably there is slip on one, or several parallel faults, because of translation of one block (or plate) past another. If the nodal planes are not parallel but either the P or the T axis is consistent in orientation, then the orientation of

that axis is probably the important parameter obtained from the analysis.

The seismic moment, dimensions of faults, and rates of deformation

A major step forward was made in seismology 15 years ago when Aki (1966, 1967) recognized that the amplitudes of long period seismic waves are proportional to the product $\mu A \bar{u}$, which he called the seismic moment, M_0 . In this expression μ is the shear modulus in the volume surrounding the earthquake, A is the rupture area, and \bar{u} is the average displacement over the faulted area. The theory behind this relationship is beyond the scope of this paper, but estimating M_0 is sufficiently easy that it has become a part of nearly all seismologists' repertoire. Uncertainties in M_0 are rarely discussed, but in general they probably are less than a factor of 2. For older events, for which the data are relatively poor, the uncertainty may be larger. Also for shallow events with one shallow dipping nodal plane, the estimated value of M_0 becomes very sensitive to the dip and increasingly uncertain for shallower and shallower dips. For strike-slip faults or for solutions with planes dipping at 30° to 60° , the uncertainty of a factor of 2 is probably a reasonable upper bound.

To estimate either of A or \bar{u} one must independently determine the other. One may estimate \bar{u} from measurements of surface displacement when surface faulting is observed. A can be estimated from (and assumed to be equal to) the area the aftershock zone. Often one can estimate a fault length, from surface faulting or from the length of the aftershock

zone. Then one may estimate \bar{u} using plausible values of the fault width. Clearly the uncertainties in all of M_0 , A and \bar{u} are comparable.

We can use the seismic moments of earthquakes to estimate the rate of deformation in a region. Brune (1968) showed that when only one fault plane is active, one can estimate the average rate of slip, v , by adding the seismic moments of earthquakes occurring in a reasonable length of time, t :

$$v = \frac{\sum_{i=1}^n M_{0i}}{\mu A t} = \sum_{i=1}^n \bar{u}_i / t$$

Here, A is the area of the entire fault under consideration. Kostrov (1974) generalized this for faults of different orientations and gave a similar expression for the strain rate in a volume (see also Anderson, 1979; Chen and Molnar, 1977). The uncertainty in the rate of slip or in the strain rate consists of four parts. One is related to the estimates of the moments and again is approximately a factor of two. A second enters with the assumed area of the fault or volume of the region. Usually we know well (\pm a few percent) the fault length or the area of the region in which we calculate the slip or strain rate, but the depth involved in deformation by earthquakes is less certain. This uncertainty (of about a factor of 2) arises both from our ignorance of the depths of earthquakes and brittle deformation and from the likely gradual brittle ductile transition. Another uncertainty arises from our ignorance of how much deformation occurs by fault creep

instead of by earthquakes, and cannot be estimated easily yet. Finally there exists the possibility that the time, t , is too short to make the present value of v a reliable estimate of long term seismic slip. Clearly Brune's and Kostrov's methods are too crude to corroborate rates of slip on deformation determined using marine magnetic anomalies and plate tectonics. Their value comes in placing constraints on the thickness of the zone of brittle deformation, on the importance of fault creep, or on the possibility that a large event is imminent. A low rate of seismic slip, compared with that obtained from plate motions, can be attributed to an unusually thin layer of brittle deformation, a large fraction of fault creep or aseismic slip, an unrepresentative seismic history, or a recent change in plate motions. Deciding which is correct requires the consideration of other data.

ACKNOWLEDGMENTS

This paper contains results from numerous studies that at the time of this writing are not yet completed. Accordingly we would like to use these acknowledgments as a platform to encourage the reader to seek out these relevant but unfinished papers. In particular, J. Armbruster, J. Baranowski, and L. Seeber did most of the work contained in Armbruster et al. (1981) on the Himalaya. A. Vilkas had not yet completed her study of the Tien Shan, and neither G. Suárez nor D. Chinn and B. Isacks had finished writing their papers on the Andes when this paper was first assembled. Moreover, we have appreciated

continuous interaction with D. Hatzfeld on the seismicity of the Atlas region and with P. Tapponnier on the tectonics of Asia in general. Finally, without a close interaction with W.F. Brace, B. Evans, and C. Goetze, the discussion of depths of foci would not have evolved to its present state. This research was supported by the National Science Foundation Grant # EAR 79-26349 and # EAR 80-18705, and NASA Grant # NAG 5-41 and # NAG 5-19.

APPENDIX A

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FIGURE CAPTIONS

Figure A1: Examples of fault plane solutions using P wave first motions of teleseismic recordings:

- (a) A well-constrained solution for a vertical strike-slip event where the nodal planes are primarily determined by the azimuthal distribution of data points.
- (b) A poorly-constrained solution for a thrust event where the position of the nodal planes are primarily determined by the distances of the data points from the center of the projection (the take-off angles). The solid curves are the preferred solution based on the observed strike of the surface rupture and the S wave polarization angles (arrows). Note the dotted curves arbitrarily drawn are also consistent with the first motions. Solid circles indicate compressional first motions (away from the source), and open circles for dilatational first motions (towards the source). The dip-slip event is an aftershock and occurred about 40 km to the south of the strike-slip event (main shock) in southern Mongolia in 1967. Data taken from Tapponnier and Molnar (1979).

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